

# Regional Moment Tensors of the 2009 L'Aquila

## Earthquake Sequence

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**Electronic Supplement:** This can be viewed at

[http://www.eas.slu.edu/Earthquake\\_Center/MECH.IT/Herrmann\\_Malagnini\\_Munafò\\_Suppl.html](http://www.eas.slu.edu/Earthquake_Center/MECH.IT/Herrmann_Malagnini_Munafò_Suppl.html)

18 **Abstract**

19

20 Broadband waveform inversion of ground velocities in the 0.02 - 0.10 Hz frequency  
21 band is successfully applied to 181 earthquakes with  $ML \geq 3$  of the April, 2009,  
22 L'Aquila, Italy, earthquake sequence. This was made possible by the development of a  
23 new regional crustal velocity model constrained by deep crustal profiles, surface-wave  
24 dispersion and teleseismic P-wave receiver functions and tested through waveform fit.  
25 Although all earthquakes exhibit normal faulting, with the fault plane dipping southwest  
26 at about  $55^\circ$  for the majority of events, a subset of events had much shallower dips. The  
27 issue of confidence in the derived parameters was investigated by applying the same  
28 inversion procedure by two groups who subjectively selected different traces for  
29 inversion. The unexpected difficulty in modeling the regional broadband waveforms of  
30 the mainshock as a point source was investigated through an extensive finite-fault  
31 modeling of broadband velocity and accelerometer data, which placed the location of  
32 major moment release up-dip and about 4-7 seconds after the initial first-arrival  
33 hypocentral parameters.

34

35

## 37 **Introduction**

38 On April 6 2009, at 01:32:39 UTC, an ML 5.8 earthquake occurred in Regione Abruzzo  
 39 (Central Italy). The initial hypocentral coordinates were 42.33°N, 13.33°E, and depth of  
 40 8.8 km (Istituto Nazionale di Geofisica e Vulcanologia web site: <http://www.ingv.it>). The  
 41 event ruptured up-dip in the southeast direction (Cirella et al, 2009), causing extensive  
 42 damage in the city of L'Aquila, and in many villages of the region. A total of 308  
 43 casualties and 1,500 injuries resulted from the collapse of buildings that could not  
 44 withstand the strong ground shaking, and 64,812 people were displaced from their homes  
 45 (Akinci and Malagnini, 2009).

46

47 The rupture occurred on the Paganica fault (Walters et al., 2009), a poorly known  
 48 structure that is now being extensively investigated (see also Emergeo Working Group,  
 49 2009). Anzidei et al. (2009) observed a maximum surface displacement of  $10 \pm 0.5$  cm  
 50 horizontally, and  $-16 \pm 2$  cm vertically, consistent with a fault plane dipping  
 51  $55^\circ \pm 2^\circ$ . Surface displacement is located on the projection of the fault plane indicated by  
 52 the spatial distribution of aftershocks (Chiaraluce et al., 2009). The best fit to the geodetic  
 53 data by Anzidei et al (2009) was achieved with a rupture surface of  $13 \times 16 \text{ km}^2$ , and an  
 54 estimated average slip of  $49 \pm 3$  cm, corresponding to an Mw 6.3 earthquake.

55

56 The strong ground motion was severe in some locations, with recorded peak accelerations

57 up to  $\sim 1g$ , clearly indicating (Akinci et al., 2010 ) the southeastward directivity of the  
58 rupture found by Cirella et al. (2009), who jointly inverted strong-motion and GPS data  
59 (Anzidei et al. 2009) for rupture properties. Atzori et al. (2009) inverted the DInSAR  
60 (Massonnet et al., 1993) co-seismic displacement for the slip distribution on the Paganica  
61 fault.

62

63 The main shock was preceded by a swarm-like activity that started a year earlier. For this  
64 study of all the events with  $ML \geq 3$ , our data set starts on October 1, 2008, at 22:47:37  
65 UTC, when an event of  $ML\ 3.1$  ( $M_w\ 3.20$ , this study) was located at  $42.59^\circ N$  and  
66  $13.29^\circ E$  (<http://iside.rm.ingv.it>) . The swarm-like activity lasted through April 6, when  
67 the main shock hit. Seven events with  $ML$  values between 3.0 and 4.0 occurred in the  
68 week preceding the main earthquake: four of them on March 30, 2009, one on April 3,  
69 and the remaining two on April 5, 2009. The entire swarm, and its abrupt acceleration in  
70 particular, may be interpreted now *a posteriori* as a precursor for the imminent  
71 occurrence of the main event. Unfortunately, it was not possible to foresee the main event  
72 before its occurrence. Four large aftershocks ( $M_w$  values 4.75, 4.81, 4.90 and 5.42, this  
73 study) occurred close to the city of L'Aquila by April 7, 2009 within 36 hours of the main  
74 shock, while another large aftershock ( $M_w\ 5.22$ , this study) occurred to the north of the  
75 city on April 9, 2009.

76

77 The occurrence of a destructive event in the vicinity of L'Aquila is not surprising, since 3  
78 large events (intensity X) affected L'Aquila in the last 650 years (1349, 1461, and 1703,

79 Stucchi et al., 2007). In recent years, some seismic sequences with a  $ML \leq 4.0$  occurred  
80 in the area, (De Luca et al., 2000; Boncio et al., 2004; Chiarabba et al., 2005; Pace et al.,  
81 2006). Deformation rates in the area were also precisely known well before the main  
82 event of April 6 2009; the area along the mountain belt is deforming in extension (2-3  
83 mm/year, Hunstad et al., 2003) within a 50 km-wide area containing the highest  
84 topographic features (Selvaggi et al., 1997). The northeast-trending orientation of the  
85 extension is consistent with focal mechanisms (Montone et al., 2004; Bagh et al., 2007),  
86 borehole breakouts (Mariucci et al., 1999) and geological data (Lavecchia et al., 1994,  
87 Westaway, 1992). Chiarabba et al. (2005) reviewed previous studies and stated that the  
88 seismotectonics along the Apennines are controlled by the north-eastward retreat of the  
89 Adria subducting slab and showed that the seismogenic layer in the region ranges  
90 between 6 and 16 km, in good agreement with the depths obtained from the waveform  
91 inversions of this study. More importantly, a number of recent studies, supported through  
92 grants of the Italian Protezione Civile (e.g., Pace et al., 2006, and Akinici et al., 2009),  
93 estimated the seismic hazard for the Central Apennines, and highlighted the elevated  
94 hazard in the area surrounding L'Aquila.

95  
96 The study by Bagh et al. (2007) investigated the background seismicity in the Abruzzo  
97 region by relocating a large number of events recorded in the previous 20 years recorded  
98 by different permanent and temporary seismic networks . They observed that the  
99 background seismicity was generally sparse with a few dense clusters due to small  
100 sequences (a few of them near l'Aquila). The seismic activity in Central Apennines, as

101 shown in Bagh et al. (2007), is distributed in the upper 15 km of the crust, and consists  
102 predominantly of normal faulting with strike parallel to the mountain belt (55% of the  
103 cases) with some pure strike-slip faulting (27% of the cases), with the remainder having  
104 trans-tensional mechanisms. Bagh et al. (2007) stated that the major active structures in  
105 the Apennines are locked normal faults, which when activated, cause secondary strike-  
106 slip structures that redistribute the perturbed stress field.

107

108 Recently INGV has upgraded the national seismic network (Amato and Mele, 2008;  
109 Michelini et al., 2008) and provided access to event recordings through ISIDE (the Italian  
110 Seismological Instrumental and parametric Data-basE which can be accessed by the link:  
111 <http://iside.rm.ingv.it/iside/standard/index.jsp> )

112 ISIDE provide access to the catalog of located earthquakes and links to the waveforms  
113 with responses. Because events are quickly posted, the event data can be quickly  
114 downloaded and processed for moment-tensor inversion. This study developed  
115 processing procedures to study the larger events of the sequence, derived a local velocity  
116 model to be used for waveform inversion, and determined that moment-tensor inversions  
117 could easily be obtained for earthquakes as small as  $ML = 3.0$ . Figure 1 shows the ISIDE  
118 location for all events with  $ML > 2$  in the vicinity of the April 6 main shock. The figure  
119 also highlights the locations of earthquakes with  $ML \geq 3$  that are the subject of this paper.

120

121 We waveform-modeled 181 events in the 0.02 - 0.10 Hz frequency band with  $ML \geq 3$   
122 that occurred in the L'Aquila region between October 1, 2008 and January 31, 2010. The

123 inversions for the moment tensor solutions were performed after we developed a regional  
124 velocity model based on profiles shown in Di Luzio et al (2009). Their Figure 5 shows a  
125 crustal geologic section taken along the CROP (CROsta Profonda) profile 11, from the  
126 Adriatic foreland on the east to the Fucino basin on the west. The CROP seismic profiles  
127 were performed in the 1980s in order to investigate the deep crust across the Apennines,  
128 and their data were recently released (Scrocca et al., 2003). The purpose of our paper is to  
129 document the inversion procedure, including the development of a regional crustal  
130 velocity model, to evaluate the capabilities of the broadband network and to understand  
131 the complex process of this earthquake sequence. We accomplish these objectives by  
132 defining the velocity model, by presenting the moment tensor solutions and then  
133 examining our difficulties in determining the source parameters of the main event of the  
134 sequence.

## 135 **Velocity Model**

136 As part of an effort for implementing routine regional moment tensor inversion in  
137 routine processing at the USGS National Earthquake Information Center, Herrmann *et*  
138 *al.* (2010) documented a procedure for systematic moment tensor inversion of continental  
139 earthquakes in the United States and Canada through a rapid grid-search procedure  
140 (Herrmann and Ammon, 2002). Much has been learned from this effort, especially as  
141 catalog completeness was extended to magnitudes less than 4.0. Signal-to-noise  
142 limitations for small earthquakes can be overcome by focusing on higher frequency  
143 content of the signal, which in turn requires velocity models capable of matching the

144 detail observed at higher frequencies. The use of the appropriate regional velocity model  
145 is important not only to match the waveforms but also to define the moment magnitude of  
146 the earthquake because the theoretical amplitudes at high frequencies depend very  
147 strongly on the velocity model.

148

149 Our preliminary processing of the L'Aquila aftershocks used a model for tectonic North  
150 America (Herrmann *et al.*, 2010) for which we had a set of pre-computed Green's  
151 functions. We quickly determined that we could perform regional moment tensor  
152 inversions using the ISIDE data sets at local magnitudes 4.0 and much lower because of  
153 the inherent high quality of the data sets and the large number of nearby broadband  
154 seismic stations. While performing quality control on the observed waveforms, we noted  
155 the presence of recognizable dispersed surface-wave trains, which suggested the  
156 application of the data processing and inversion tools of Herrmann and Ammon (2002) to  
157 define a specific velocity model for use in the study area.

158

159 We made group velocity measurements using multiple filter analysis (Herrmann, 1973)  
160 on 80 vertical and transverse component waveforms for 6 aftershocks to yield about 600  
161 Rayleigh- and Love-wave dispersion measurements in the 4.4 to 28 second period range,  
162 being careful not to select the longer periods at short epicentral distances for which the  
163 dispersion was not yet well developed. The aftershocks and stations used for the group  
164 velocity study, Figure 2, sample the central Apennines, and thus any derived velocity  
165 model is appropriate for these paths or for a similar structural environment.



166

167 The starting model, given in the Appendix, was based on the work of Di Luzio *et al*  
168 (2009) who interpreted the seismic data from a deep seismic reflection profile across the  
169 Apennines that passed near L'Aquila. The crustal model for their stations 7-8, near  
170 L'Aquila, was used to define the deeper crustal boundaries and P-wave velocities. The  
171 surface wave inversion program, *surf96* (Herrmann and Ammon, 2002), was run with a  
172 smoothing constraint to find a simple model that matches the observations. To be  
173 consistent with the major structural boundaries in the work of Di Luzio *et al.* (2009), we  
174 applied stronger weighting to permit a basin boundary at a depth of 3 km, and fixed the  
175 velocities of the halfspace and deepest crustal layer in the model. We permitted the other  
176 crustal velocities, with emphasis on the upper crustal velocities, to change since the  
177 surface-waves are the dominant signal for the time-domain moment tensor inversion and  
178 are in turn affected by upper crustal S-wave velocities. Moreover, the strong P-wave  
179 signal often observed out to 100 km also is controlled by the upper crustal velocities.  
180 The starting model has a low-velocity in the mid-crust because of the westward  
181 subduction beneath the Apennines (see Chiarabba et al, 2005). The resulting surface-  
182 wave based velocity model given in the Appendix, named CIA (Central Italian  
183 Apennines), is thus constructed to be consistent with earlier studies as well as the  
184 measured dispersion.

185

186 Being aware that fundamental mode surface-wave dispersion data cannot resolve sharp  
187 discontinuities in the velocity model, we also assembled a representative data set of

188 radial-component P-wave receiver functions for the MedNet station AQU at L'Aquila for  
189 9 earthquakes using the low-pass filter parameter  $\alpha = 1.0$  with the time-domain iterative  
190 deconvolution technique of Ligorría and Ammon (1999). The station AQU was selected  
191 for analysis because it lies within the region for which the velocity model is required and  
192 since waveforms were easily available from data archives. Since many crustal studies  
193 make use of receiver functions, neglecting their use would call into question the value of  
194 a velocity model based only on surface waves. These receiver functions were inverted  
195 together with the dispersion data using the program *joint96* (Herrmann and Ammon,  
196 2002) to yield the joint surface-wave dispersion – receiver function model given in the  
197 Appendix as ACI (Appennino Centrale d'Italia). Since our objective was to augment the  
198 CIA model determined using *surf96*, the CIA model was used as the starting model, with  
199 the difference that we subdivided many layers to be able to fit the finer features of the  
200 receiver functions. We did not permit the half-space velocity to change and again placed  
201 more emphasis on the change in layer velocities in the upper 10 km because of the  
202 ringing character of the receiver functions is strongly affected by the presents of low  
203 velocity sedimentary basins.

204

205 Figure 3 compares our observed dispersion with the predictions of the CIA, ACI, BAGH  
206 (Bagh et al, 2007) and TDMT (Scognamiglio et al, 2009) models. The scatter in the  
207 observed dispersion is related to location and origin time error, the effect of 3-D  
208 structure, and biases in the multiple-filter analysis determinations. However, the mean is  
209 assumed stable enough to define the dispersed shape of the observed waveforms. The

upper 1.5 km of the Bagh et al. (2007) model was modified to have lower velocities in accordance with borehole information in the L'Aquila region (pers. comm. L. Scognamiglio, 2009). The TDMT model is used for the INGV regional moment tensor determination. The TDMT model cannot match the observed dispersion because of the thick low velocity layers near the surface that give rise to the very low fundamental mode group velocities at shorter periods. The BAGH model is better at shorter periods, but our *ad hoc* extension of the model to depths greater than the 20 km of the Bagh et al (2007) model was not adequate and demonstrates the need for defining the complete crustal model. Since both the CIA and ACI models were based on the inversion of the dispersion data, they fit the observed dispersion well.

220

Figure 4 shows the result of the joint inversion of the surface-wave dispersion and P-wave receiver functions at AQU. The figure shows both the starting and final models for the inversion, CIA and ACI, respectively. Although the receiver function fit is not perfect, the observed ringing has begun to be fit. For this station the ringing, due to the effect of the shallow velocity structure, dominates any effect of deeper crustal structure beneath the MedNet station L'Aquila

227

Figure 5 compares the four models. The low velocities of the upper 8 km of the TDMT are obvious, as is the assumed higher velocity lower crust of the BAGH model. The additional detail in the ACI model (solid gray line) compared to the simpler CIA model (solid black line) is required to model the observed long duration ringing of the AQU

232 receiver functions.

233

234 For use in source inversion, we initially computed Green's functions for both the CIA  
235 and ACI models, and found they were similar when these were filtered in the 0.02 –  
236 0.10 Hz band used for the source inversion, which is not surprising since both fit the  
237 observed dispersion in the same way. For reasons of computational speed, we used the  
238 simpler CIA model to compute an extensive set of Green's functions for depths between 1  
239 and 29 km in 1 km increments, and epicentral distances between 1 and 350 km at 1 km  
240 increments. A perfectly elastic model is used since the effect of reasonable Q values in  
241 modeling observations in the low frequency band and at the short epicentral distances  
242 would be negligible.

243

## 244 **Moment Tensor Solutions**

245

246 When we concluded that it was possible to determine source parameters for earthquakes  
247 with  $ML \geq 3$ , we developed bash shell scripts to ensure a uniform approach to the  
248 inversion and to reduce the need for manual intervention. The event location from ISIDE  
249 was used to initiate the processing. The ISIDE archive containing the waveforms and  
250 corresponding pole-zero files was unpacked. An initial QC (quality control step), applied  
251 to eliminate waveforms with data gaps or noisy signals, was followed by a second QC  
252 that examined the deconvolved ground velocity waveforms in the 0.02 – 0.10 Hz band

253 typically used for inversion. A final QC of the inversion results served to identify  
254 problematic waveforms, which were eliminated. A web page presenting a record of all  
255 processing steps can be viewed at [http://www.eas.slu.edu/Earthquake\\_Center/MECH.IT/](http://www.eas.slu.edu/Earthquake_Center/MECH.IT/).  
256 In spite of trans-Atlantic download times, we often had a solution posted on the web  
257 page within 30 – 60 minutes of the event notification.

258

259 The grid search for source parameters uses filtered ground velocity as a data set to search  
260 for the best fitting shear dislocation characterized by the strike and dip of the fault plane  
261 and the rake angle giving the direction of fault movement on the fault plane. For each  
262 source depth, a search is performed over all values of strike, dip and rake angles at 10°  
263 increments, followed by a finer 5° search in a region  $\pm 20^\circ$  about the crude best fit. The  
264 best fit is defined as the greatest reduction in weighted variance with each trace weighted  
265 as a function of epicentral distance in a manner that is proportional to distance out to  
266 100 km and inversely proportional to distance beyond 100 km to overcome the  
267 dominance of large amplitudes and the effects of mis-location on azimuth at short  
268 distance, and inadequacies in the velocity model at larger distances. The Herrmann et al  
269 (2010) grid search algorithm permits a time shift to better align the waveforms to  
270 overcome mis-location and slight inadequacies of the Green's functions for the path to  
271 each station. We have found that the derived time shift is diagnostic of mis-location error  
272 and the need for velocity model improvement.

273

274 The determination of the passband for inversion is critical. We accomplish this by

275 applying a 3-pole causal high-pass Butterworth filter at the lower corner followed by a 3-  
276 pole causal low-pass Butterworth filter at the upper corner. The corner frequencies were  
277 selected on the basis of the expected fundamental mode surface-wave spectral shape, on  
278 avoiding instrumental and ground noise at low frequencies, microseism noise, and the  
279 consequence of using an imperfect crustal model at higher frequencies. The upper corner  
280 should also be adjusted to be lower than the corner frequency of the earthquake – the 0.02  
281 – 0.10 Hz band was used for all earthquakes except for the main shock for which we used  
282 the 0.01 – 0.025 Hz frequency band. The choice of using ground velocity filtered in the  
283 0.02 - 0.10 Hz frequency band was made to be able to analyze small earthquakes and to  
284 check on the appropriateness of the velocity model in as wide a bandwidth as possible.

285

286 Of the 235 earthquakes in the INGV catalog in the 2008/10/02 – 2010/01/31 time period  
287 with  $ML \geq 3$  and greater, we were able to determine source parameters for 181 of these  
288 earthquakes. As an example of the processing, consider the event of 2009/08/12 14:51:33  
289 UTC. For this earthquake we selected 23 vertical-component (Z), 7 radial-component  
290 (R), and 10 transverse-component (T) waveforms for inversion. Figure 6 shows the  
291 stations used in relation to the epicenter. The epicentral distances range from 18 to 146  
292 km. Figure 7 plots the goodness of fit, the reduction in distance weighted variance as a  
293 function of source depth, with the best mechanisms associated with each source depth;  
294 the best fit occurs at a depth of 7 km. Figure 8 overlays the filtered observed waveforms  
295 on top of those predicted for the best solution. There is an excellent fit between observed

296 and predicted waveforms. More importantly, the time shifts, indicating the shift of the  
297 predicted (dashed) with respect to the observed (solid) traces, are typically on the order  
298 of 1sec, which indicates a consistency of the ISIDE source location and origin time (our  
299 relocation using the CIA model with our arrival time picks gave the same epicenter and  
300 origin time) as well as the applicability of the CIA velocity model. The increasingly  
301 negative time shifts for the Rayleigh-wave pulse on the Z component as distance  
302 increases, indicates that the model could be about 3-4% faster for the Rayleigh waves.  
303 The shapes of the observed and predicted signals match well.

304

305 To address the fundamental issue of the usefulness of this model, we ran the inversion in  
306 different frequency bands with the results shown in Table 1. Although the goodness of fit  
307 depends on the frequency band used, the source parameters are quite similar. At the  
308 lowest frequency, the reduced fit parameter is due to long period noise. At the highest  
309 frequency, the effect of scattered waves degrades the fit.

310

311

312

313 Figure 9 compares the observed and predicted waveforms for the station FDMO which is  
314 at an epicentral distance of 78.7 km. The observed signal shapes and peak amplitudes are  
315 fit well by the synthetics for the best solution in each frequency band. The well-  
316 developed surface wave dispersion that led to the development of the CIA model is

317 obvious. The time shifts indicate that small modifications to make the Rayleigh wave  
318 slightly faster and the Love wave slightly slower could be made to the velocity model,  
319 but the source solution would not change significantly. The obvious presence of the  
320 surface wave, even for this 7 km depth earthquake, has implications for ground motion  
321 scaling at periods as short as 2 seconds, demanding the use of surface-wave rather than S-  
322 wave scaling with distance.

323

324 Figure 10 summarizes the completeness the source parameter catalog that we were able to  
325 compile. Only for  $ML < 3.3$  is there any significant lack of completeness. Most of the  
326 missing small earthquakes occurred within the first day of the aftershock sequence, when  
327 their low frequency content was buried in the incoherent low-frequency coda of the main  
328 shock. Figure 11 compares our moment magnitudes to the automatic network magnitudes  
329 – there is a very good correspondence. However, the moment tensor inversion depths do  
330 not correlate with the automatic depth determination of the network; this is not surprising  
331 given the dependence of depth on the assumed velocity model and on the distribution of  
332 the permanent network stations. Chiaraluce et al (2010) recomputed source depths by  
333 carefully re-reading arrival times and by using a linear gradient velocity model. A  
334 comparison of our depths to theirs shows a better correlation. This latter comparison is  
335 not sufficient to demonstrate the correctness of our source depths because we use  
336 different velocity models, but we argue that fitting the waveform with a calibrated local  
337 velocity model, especially the large surface-wave, provides a much stronger constraint on



338 source depth than using only first-arrival data.

339

340 Figure 12 summarizes the source parameters contained in our catalog (electronic  
341 supplement). The lower hemisphere focal mechanism plots indicate a normal-faulting  
342 environment with the tension axis normal to the trend of the Apennines. Excluding the 11  
343 events which having nodal plane dips  $< 25^\circ$  to the southwest, the mean dip is  $57^\circ$  to  
344 the southwest with a standard deviation of  $13^\circ$ . Chiaraluce et al (2010) noted that some  
345 of the 11 solutions with the shallow dips correlated well with a flattening of hypocenters  
346 with depth in the northern part of the study area. We also note, as have others, that the  
347 pattern of earthquakes with  $M > 3$  shows three groups of hypocenters.

348

349 Source inversion was performed independently by the SLU and INGV authors using the  
350 same Green's functions and inversion code. The only difference affecting results is the  
351 subjective choice of waveforms used for the inversion. Having two catalogs permits an  
352 analysis of the variability in the source parameter estimates, which is summarized in  
353 Table 2. This table lists the variability in the source depth,  $H$ , the moment magnitude and  
354 the strike, dip and rake angles. Care was taken to compare similarly oriented nodal  
355 planes and the ambiguity of using angles, e.g., strikes of  $0^\circ$  and  $360^\circ$  are equivalent as are  
356 rakes of  $+180^\circ$  and  $-180^\circ$ . There were more outliers in the strike and dip values than in  
357 the  $H$ ,  $M_w$  and Dip, but the variability was roughly Gaussian. To avoid any possible bias  
358 in the angles and since the earthquakes all represent normal faulting, we also looked at  
359 the angles between the P-axis vectors for each strike, dip and rake combination, and

360 similarly the angles between the T-axis and the null B-axis. These angles vary between  
361  $0^\circ$  and  $90^\circ$ , and exhibit an approximately Poisson distribution. The entries in this table  
362 serve as a guide to confidence in this type of source parameter estimate. Scognamiglio et  
363 al (2010) used the CIA velocity model with a different source inversion code to  
364 determine the parameters of all earthquakes with  $M_L > 3.5$ . A cursory comparison of our  
365 moment magnitudes and source depths to theirs indicates that the confidence values in  
366 Table 2 are acceptable.

367

## 368 **L'Aquila Mainshock**

369

370 Figure 12 also highlights the fact that the main shock of the L'Aquila sequence, the  
371 earthquake of 2009/04/06 01:32:39 appears to be shallower than adjacent aftershocks.  
372 We initially had difficulty determining the source parameters of the main shock. To avoid  
373 having to worry about corner frequency effects for a large earthquake, we initially used  
374 the 0.01 – 0.05 Hz frequency band for the inversion, which was the appropriate choice for  
375 the 2008/02/21  $M_w=5.88$  Wells, Nevada, earthquake  
376 ([http://www.eas.slu.edu/Earthquake\\_Center/MECH.NA/20080221141605/](http://www.eas.slu.edu/Earthquake_Center/MECH.NA/20080221141605/)). Figure 13  
377 shows the goodness of fit with depth corresponding to this choice – the lack of sensitivity  
378 to depth and the tendency toward a large source depth was not satisfying, especially since  
379 the source inversions of the aftershocks led to much shallower depths. We then used the  
380 0.01 – 0.025 Hz frequency band (D. Dreger, personal communication) which led to

381 Figure 14 and a shallower depth estimate of 5 km. Finally we added more distant stations  
382 (L. Scognamiglio, personal communication) to reduce sensitivity on nearby, perhaps  
383 overdriven sensors on the estimate of the moment magnitude. Figure 15 shows the  
384 locations of the stations used for the final broadband inversions, and for the sensitivity  
385 studies to follow. The epicentral distances vary from 51 to 414 km; 7 stations are at  
386 distances less than the 146 km used in the model validation study while 13 are at greater  
387 distances.

388

389 Figure 16 compares the observed and predicted waveforms for the best solution using the  
390 0.01 – 0.025 Hz passband. The goodness of fit parameter was 0.698 when using a time  
391 window 10 seconds before and 180 seconds after the predicted P-wave first arrival time.  
392 When using the time window from 0 – 250 s after origin time, the best fit was 0.695,  
393 which is very similar because of the high signal-to-noise ratio. The time shifts in Figure  
394 16 are uniformly positive, but seem to be path-dependent with smaller Z-component time  
395 shifts to the northwest of the mainshock, perhaps an indication of the need for path-  
396 dependent models. The time shifts are much larger than required for the many aftershock  
397 solutions.

398

399 The requirement for large time shifts to match waveforms can be due to the use of the  
400 wrong velocity model for the Green's functions, hypocenter error, or a distributed source  
401 process. We discount the model problems for the stations at short distance because of the

validation study of Figure 8. Table 3 is a tabulation of locations available for the main shock. This tabulation consists of the initial and final locations on ISIDE, our relocations using our P-wave first arrival picks with the Computer Programs in Seismology location code, *elocate* (Herrmann and Ammon, 2004), together with our CIA model, the Michelini et al (2009) relocation that considered two velocity models, and the Chiaraluce et al (2010) location using re-picks of all data and a gradient model. Our locations using *elocate* used 24 P-wave picks from the broadband stations (BB), 31 P-wave picks from the Italian Accelerometric Archive – ITACA ( <http://itaca.mi.ingv.it/ItacaNet/>) (ACCEL) and the combined set of 55 phases. All relocations are moved a few kilometers east with origin times about 1 second later than the initial ISIDE location. Assuming a 2.5 – 3.0 km/s group velocity for the surface waves, these slight differences in the position of the hypocenter cannot explain the large time delays on the order of 5 seconds seen in the point source inversion results in Figure 16. Although the depths are deeper than the 5 km obtained from the point source inversion, one can argue that the source inversion is sensitive to the centroid of moment release, and the eastern shift of the hypocenter moves the main shock into a zone of shallower aftershock depths. The use of the CIA model, which has much lower velocities in the upper 1.5 km than even the LI07 model used by Michelini et al (2009) yields the deepest source depth estimate, however the difference in depth will not significantly affect the surface-wave timing, which make up the largest part of the signal. We concluded that the source of significant moment release must be shallower and later than the *elocate* solution.

423

424

425 We then decided to apply a simple finite source inversion to fit the regional broadband  
426 waveforms by adapting the work of Hartzell and Heaton (1983). We defined a  
427 rectangular fault grid, used Green's functions appropriate to the center of each grid-cell  
428 to the nearest kilometer in epicentral distance and source depth, and let the rupture start at  
429 the hypocenter and propagate with a velocity at a fixed fraction of the local S-wave  
430 velocity obtained from the CIA model. We assumed that the rise-time was fixed at 1.0  
431 second, a value selected to avoid Gibb's phenomena in the Green's functions, which were  
432 computed with a sample interval of 0.25 sec. Since the observed waveforms will be  
433 modeled in the 0.01 – 0.025 and 0.01 – 0.05 bands, the effect of any reasonable  
434 subevent rise-time will not be resolvable. To permit comparisons with the point source  
435 moment tensor solutions, the same distance weighting function is applied for the finite  
436 fault inversion and for the final characterization of goodness of fit or reduction of  
437 variance. Although we also investigated different rupture initiation points and different  
438 rupture velocities, we present the results for just one hypocenter and rupture velocity  
439 since our objective is not to provide the definitive mapping of moment release and slip on  
440 the fault, but rather to understand both the need for the low frequency passband and the  
441 source of the large time shifts for the point source solution. The comparisons entail  
442 fitting the 250-sec ground velocity window following the origin time. In all cases the  
443 CIA model is used for the Green's functions.

444

445 The fault is characterized by 20 segments of length 2 km along strike, and 25 segments of  
446 width 1 km down dip. The size of the fault plane was chosen to encompass the zone of  
447 initial aftershocks and the hypocenter was centered horizontally in the strike direction.  
448 The hypocenter is on the fault at a depth of 13 km at coordinates 42.339°N and 13.371°E,  
449 the fault strikes at 135°, dips at 55° and has a fixed rake of -95°, which are the parameters  
450 determined by the grid search for the main shock. We chose this nodal plane since the  
451 moment tensor solutions of the aftershocks indicate a trend of increasing depth to the  
452 southwest. The total moment release is fixed at  $M_w=6.13$ . The system of equations to be  
453 solved is

$$454 \quad [\alpha A, U, \gamma S]^T M = [\alpha d, M_0, 0]^T.$$

455 Here  $M$  is an  $n \times 1$  matrix giving the seismic moment release in each of the  $n$  cells.  $A$  is  
456 an  $m \times n$  matrix of predicted waveforms for each cell,  $U$  is a  $1 \times n$  matrix of 1's,  $S$  is an  $n \times n$   
457 Laplacian smoothing matrix,  $d$  is a matrix of the waveforms to be fit and  $M_0$  is the fixed  
458 seismic moment. The scaling factor  $\alpha$  is selected so that the row-norm of  $A$  is unity. The  
459 factor  $\gamma$  controls the degree of spatial smoothing. Table 4 compares the goodness of fit for  
460 the point source and finite fault solutions in different passbands as a function of the  
461 spatial smoothing factor, the data sets and rupture velocity. In the 0.01 – 0.025 Hz band  
462 the fits are essentially the same, although the finite fault simulation defines the moment  
463 release in cells such that the time shift problem is addressed. This similarity in fits may be  
464 an indication that the point source solution is adequate in the 0.01 – 0.025 Hz passband.

465 In the 0.01 – 0.05 Hz passband, the finite fault fit is better than for the point source  
466 solution. In the 0.01 - 0.05 Hz passband, the fit parameters are lower, because this  
467 inversion technique did not permit small time shifts that account for path-dependent  
468 propagation differences.

469

470

471

472 Figure 17 compares observed and predicted waveforms for the 0.01 – 0.025 Hz passband  
473 and the finite fault parameters of Table 4. The waveforms are well fit in time, except for  
474 the surface wave signal on the transverse component at large distances, indicating the  
475 need for a slight refinement in the CIA model. Figure 18 shows the derived discrete  
476 finite-fault sources in the 0.01 – 0.025 Hz passband relative to the locations of the  
477 moment tensors that we determined in this study. The hypocenter used for the finite fault  
478 inversion is indicated by the star and the finite fault events by the diamonds. The two  
479 largest sub-events have moment magnitudes of 5.7 and 5.6. Similar plots for inversions in  
480 the 0.01 – 0.05 Hz passband yielded essentially the same pattern. The fact that the  
481 moment release is 3.5 – 6 seconds after the assumed origin time and that the subevents  
482 are distributed  $\pm 10$  km along strike with respect to the position of the hypocenter, act to  
483 explain the time shifts required in the point source solution of Figure 16. We also note  
484 that 90% of the moment release is at depths less than 6 km, in agreement with the point  
485 source depth estimate and that few aftershocks are in the region of major moment release.

486

487 Although the solution shown in Figure 18 provides a very good fit to the observed  
488 waveforms, it is not realistic in that the corresponding slips are excessive because of the  
489 small cell areas. Rather than plotting the seismic moments of the sub-events, Figure 19  
490 presents the derived slip,  $u = M_0/(\mu A)$  where  $\mu$  and  $A$  are the rigidity and area of each  
491 cell, respectively. The figure shows the effect of smoothing, which spreads out the  
492 distribution of slip on the fault. The value of the slip has a tendency to be larger at  
493 shallower depths, because of the smaller rigidities. The common feature of these three  
494 inversions is that the fault slip is in the upper 6-7 km and that the time of major slip is  
495 delayed 3.5 to 7 seconds after the initial break at depth.

496

497 Although this numerical exercise accounts for the time shifts required by the point  
498 source solution, the sensitivity of the solution to rupture velocity and the usefulness of the  
499 distant broadband data set must be addressed. We combined the ZNE component  
500 accelerogram data from ITACA, integrated to velocity, with the ZRT broadband data  
501 and inverted the entire data set in the 0.01 – 0.05 Hz passband, the same passband that  
502 could not be used to characterize the mainshock as a point source. The locations of the  
503 accelerographs are indicated in Figure 15 by the inverted triangles. Specifically we added  
504 the stations ANT, AQA, AQU, AQV, ASS, AVZ, BOJ, CDS, CHT, CLM, CMB, CMR,  
505 CS01, CSS, FMG, GSA, IRS, LSS, MMP, MTR, ORC, PTF, SBC, SPC, SPO, SUL,  
506 TMO and VRP which ranged in epicentral distance from about 3 to 140 km. In general,  
507 the fits improved with acceleration data included, because of the addition of the simpler



508 waveforms at short distances.

509

510 Figure 20 shows the derived slip on the fault plane as a function of assumed rupture  
511 velocity as a fraction of the local medium S-wave velocity for the two data sets with the  
512 smoothing parameter  $\gamma = 1$ . In order to fit the signal delays seen in Figure 16, the  
513 position of maximum slip becomes shallower as the rupture velocity increases because  
514 the inversion is in absolute time. We also see that the magnitude of maximum slip  
515 increases with increasing rupture velocity because more of the moment release is at  
516 shallow depths. The shapes and locations of the major slip are similar for both waveform  
517 data sets.

518

519 Figure 21 decreases the smoothing parameter to  $\gamma = 0.1$ , with the consequence that more  
520 character is seen in the slip distributions. Again there is similarity in the patterns derived  
521 from the two data sets for the same rupture velocity parameter. However the addition of  
522 acceleration data sharpens the slip pattern. The goodness of fit associated with these  
523 inversions are all better than that of the point source solution for the 0.01 – 0.05 Hz  
524 frequency band.

525

## 526 **Discussion**

527

528 This study was able to provide a very complete moment tensor catalog of 181

529 earthquakes down to  $M_w=3$  for the L'Aquila earthquake sequence for several reasons: the  
530 Istituto Nazionale di Geofisica e Vulcanologia has a dense broadband digital seismic  
531 network in the epicentral region, the earthquakes occurred at the time of year when  
532 microseismic noise started to decrease, the earthquakes generated high amplitude  
533 dispersed surface waves because of the local velocity structure at shallow depths, and  
534 finally, the aftershock sequence was very energetic in terms of the numbers of  
535 aftershocks with  $M_w \geq 3.0$ . Normal faulting with almost all tension axes in the E to ENE  
536 directions characterizes the solutions.

537

538 Our catalog of regional moment tensor solutions differs very little from that developed by  
539 Scognamiglio et al (2010) since they use the same CIA velocity model and similar filtered  
540 ground velocity waveforms. Details of the small differences in the two catalogs are given  
541 in their paper. Their effort, though, focused on earthquakes with  $M_w \geq 3.5$  and on  
542 automatic processing. A comparison of 25 regional centroid moment tensor solutions  
543 determined by Pondrelli et al (2010) for the larger earthquakes showed that our moment  
544 magnitudes were smaller by 0.22  $M_w$  units and our depths were shallower by 5 km than  
545 theirs. We attribute this difference to our use of waveform data within 200-300 km to the  
546 exclusion of any paths through the sea, the use of high frequencies and, more importantly,  
547 a crustal model calibrated for the propagation paths used. In simple terms, the moment  
548 magnitude value is not independent of the velocity model used, which must be presented  
549 alongside the  $M_w$ 's.

550

551 We found that the determination of the source parameters of the main shock required  
552 much care in the selection of the frequency band and data sets for inversion. Although we  
553 initially assumed that the 0.01 – 0.05 Hz band would be adequate, given previous  
554 experience with the similar sized 2008 Wells, Nevada, normal faulting event, the grid  
555 search solution diverged to depths deeper than expected for the source region. The use of  
556 the lower 0.01 – 0.025 Hz passband, at the suggestion of D. Dreger, alleviated the  
557 problem, but the goodness of fit did not show as well defined sensitivity to depth as seen  
558 in application of the same procedures to the smaller aftershocks (e.g. Figure 7). We are  
559 not sure how much of this problem arises from the non-uniform station distribution in  
560 azimuth because of the geometry of the Italian Peninsula or because of the lack of data at  
561 short distances because of overdriven sensors. From experience in inverting surface-  
562 wave spectral amplitudes (Herrmann et al, 2010), we know that the effect of increasing  
563 depth is to reduce the high frequency content of the fundamental mode surface waves  
564 (Tsai and Aki, 1971). This effect must also be apparent in the time domain if the  
565 surface-wave is the dominant part of the observed signal. Our finite fault inversions  
566 yielded a sequence of shallow events distributed in time and space, which have the effect  
567 of modifying the higher frequency content of the observed signal due to signal  
568 interference at high frequencies, when compared to that of a point source. The point  
569 source inversion interprets this effect as an increased source depth. The important lesson  
570 learned is that if the goodness of fit, as seen in Figure 13, is observed in a region where  
571 one expects upper crustal earthquakes, then one should invert the data again using a lower

572 passband and also consider the source to be spatially complex.

573

574 A second indication of a spatially complex source was the fact that large time shifts were  
575 required to match observed waveforms data, even when using the lower frequency band.

576 We investigated the effect of different data sets and velocity models, and concluded that  
577 the epicenter was not very sensitive to the velocity model or location technique and that  
578 the differences in origin time were not sufficient to explain the required time shifts in the  
579 inversion of the regional broadband data sets. We concluded that the centroid of moment  
580 release was not the hypocenter based on first arrivals. This simple finite source  
581 inversion demonstrated the ability to fit the regional waveforms in absolute time because  
582 of the use of a calibrated regional velocity model.

583

584 Our finite fault modeling of regional broadband waveforms requires that the major  
585 moment/slip release to occur roughly 4-7 seconds after the origin time and up-dip of the  
586 hypocenter. The use of regional and local data sets in the 0.01 – 0.05 Hz passband cannot  
587 resolve issues of the choice of rupture velocity and degree of spatial smoothing, other  
588 than that smoothing is required to avoid extreme values of slip and that the distance of  
589 the major fault slip is a function of the rupture velocity. We also note that the goodness  
590 of fit parameter cannot be used as the sole criteria for defining the solution since, as we  
591 have seen, physically unrealistic answers of large slip may result. The resolution of  
592 these questions cannot be accomplished without other data, such as measurements of

593 permanent deformation near the fault from integrated accelerograms, GPS or InSar, or  
594 perhaps from broadband modeling of teleseismic data which will be sensitive to the depth  
595 of the slip release.

596

597 The shallow depths of major slip is comparable to that estimated by Atzori et al (2009)  
598 from an inversion of DInSar data. It is also interesting that there are few significant  
599 (e.g.,  $M_L > 3$ ) aftershocks associated with the unsmoothed inversion of low frequency  
600 data shown in Figure 18. Neither the Atzori et al (2009) nor any of our solutions are  
601 compatible with the inversion of GPS and strong motion data by Cirella et al (2009) who  
602 have the major slip at about  $42.28^\circ\text{N}$  and  $13.43^\circ\text{E}$ , near the location of the large, deep  
603 aftershock seen in Figure 18. However a reevaluation of the inversion of GPS and strong  
604 motion data using the CIA velocity model developed in this paper (Scognamiglio et al  
605 2010) has major slip up-dip from the hypocenter with with directivity to the southeast.

606

607 The inversion of just the broadband data did serve to highlight the spatial location of the  
608 shallow moment and slip release in a manner that overcame the initial bias due to the  
609 first-motion hypocenter by moving the large fault motions up-dip from about 13 km to 5  
610 km deep, a significant change in terms of expected surface motions.

611

612 The L'Aquila main shock is interesting for another reason. What is the significance of the  
613 initial hypocenter to the main moment release? If our finite fault solution that fits regional

614 waveforms in travel time is reasonable, is the initial event at depth the trigger for the  
615 large earthquake, or is it just coincidental? The first option seems more reasonable from  
616 the point of view of modeling the regional broadband waveforms because it is hard to  
617 conceive of a near instantaneous rupture process that propagates horizontally, while the  
618 shear-waves from a trigger earthquake at depth might reach the shallow slip regions in  
619 the appropriate sequence.

620

621 This study provides a unique compilation of waveform constrained source parameters for  
622 over 180 aftershocks. The depths and source mechanisms are constrained well by the data  
623 sets, and have already been used in the interpretation of the dynamics of the sequence and  
624 the identification of fault structures that were activated by the changes in stress during  
625 this sequence.

626

627 Because of our success in deriving source parameters for small events, we have extended  
628 the application of our inversion technique to the entire peninsula, not only because of  
629 curiosity about the source process but also as a test of the spatial limits on the  
630 applicability of our regional crustal model. Having one or more regional velocity  
631 models with pre-computed Green's functions available is an essential part of being able  
632 to automate the moment tensor determination in order to be prepared for the next large  
633 earthquake in Italy.

634

## 635 DATA AND RESOURCES

636

637 Some figures were created using the GMT package of Wessel and Smith (1991).

638 Broadband waveform data from the Italian Seismological Instrumental and Parametric

639 Data-Base (ISIDE) is available at the URL (last accessed October 20, 2010)

640 <http://iside.rm.ingv.it/iside/standard/index.jsp>

641 The strong motion data from the Italian Accelerometric Archive (Itaca) is available at

642 URL (last accessed October 20, 2010)

643 <http://itaca.mi.ingv.it/ItacaNet/>

644 Computer Programs in Seismology is available at (last accessed October 20, 2010)

645 [http://www.eas.slu.edu/Earthquake\\_Center/CPS/CPS330.html](http://www.eas.slu.edu/Earthquake_Center/CPS/CPS330.html)

646

## 647 ACKNOWLEDGMENTS

648

649 L. Malagnini was supported under the auspices of the Dipartimento della Protezione

650 Civile, under contract S3-INGV-DPC (2007-2009), project Valutazione rapida dei

651 parametri e degli effetti dei forti terremoti in Italia e nel Mediterraneo. The work of R.

652 Herrmann was supported in part by USGS IPA 6-8931-2458. We wish to thank D. Dreger

653 and L. Scognamiglio for their assistance in the selection of frequency bands and

654 broadband waveforms for the analysis of the mainshock. C. Ammon, K. Koper and M.

655 Woods provided welcome suggestions for the improvement of this paper.

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## 663 **References**

664 Akinci, A., and L. Malagnini (2009). The 2009 Abruzzo earthquake, Italy, *IRIS*  
665 *Newsletter*, Year 2009, Issue 1.

666 Akinci, A., L. Malagnini, and F. Sabetta (2010). Characteristics of the Strong Ground  
667 Motions from the 6 April 2009 L'Aquila earthquake, Italy, *Soil Dynamics And*  
668 *Earthquake Engineering*, 30, 320-335, doi:10.1016/j.soildyn.2009.12.006.

669 Akinci, A., F. Galadini, D. Pantosti, M. Petersen, L. Malagnini, and D. Perkins (2009).  
670 Effect of time-dependence on probabilistic seismic hazard maps and deaggregation  
671 for the central Apennines, Italy, *Bull. Seism. Soc.Am.*, 99, 2A, 585-610,  
672 doi:10.1785/0120080053.

673 Amato, A., and F. Mele (2008). Performance of the UNGV National Seismic Network  
674 from 1997 to 2007, *Annals of Geophysics* 51, 417-431.

675 Anzidei, M., E. Boschi, V. Cannelli, R. Devoti, A. Esposito, A. Galvani, D. Melini, G.  
676 Pietrantonio, F. Figuzzi, V. Sepe and E. Serpelloni (2009).. Coseismic deformation of



677 the destructive April 6, 2009 L'Aquila earthquake (central Italy) from GPS data,  
 678 *Geophys. Res. Lett.*, **36**, L17307, doi:10.1029/2009GL039145.

679 Atzori, S., I. Hunstad, M. Chini, S. Salvi, C. Tolomei, C. Bignami, S., Stramondo, E.,  
 680 Trasetti, A. Antonioli, and E. Boschi (2009). Finite Fault Inversion of DinSAR  
 681 coseismic displacement of the 2009 L'aquila earthquake (central Italy), *Geophys. Res.*  
 682 *Lett.*, **36**, L15305, doi:10.1029/2009GL039293.

683 Bagh, S., L. Chiaraluce, P. De Gori, M. Moretti, A. Govoni, C. Chiarabba, P. Di  
 684 Bartolomeo, M. Romanelli (2007). Background seismicity in the Central Apennines  
 685 of Italy: The Abruzzo region case study, *Tectonophysics*, **444** 80 – 92,  
 686 doi:10.1016/j.tecto.2007.08.009

687 Boncio, P., G. Lavecchia, G. Milana and B. Rozzi (2004). Improving the knowledge on  
 688 the seismogenesis of the Amatrice–Campotosto area (central Italy) through an  
 689 integrated analysis of minor earthquake sequences and structural data. *Ann. Geophys.*  
 690 **47**, 1723–1742.

691 Chiarabba, C., L. Jovane and R. DiStefano (2005). A new view of Italian seismicity using  
 692 20 years of instrumental recordings. *Tectonophysics* **395**, 251–268. subsurface data.  
 693 *Tectonophysics* **317**, 273–297.

694 Chiaraluce L., C. Chiarabba, P. De Gori, R. Di Stefano, L. Improta, D. Piccinini, A.  
 695 Schlagenhauf, P. Traversa, L. Valoroso and C. Voisin. The April 2009 L'Aquila  
 696 (Central Italy) Seismic Sequence. Submitted to *Bull. of Theor. and Appl.*  
 697 *Geophysics*. [HAS THIS BEEN PUBLISHED YET?]

698 Cirella , A., A. Piatanesi, M. Cocco, E. Tinti, L. Scognamiglio, A. Michelini, A. Lomax,

699 and E. Boschi (2009). Rupture history of the 2009 L'Aquila (Italy) earthquake from  
 700 non-linear joint inversion of strong motion and GPS data, *Geophys. Res. Lett.*, 36,  
 701 L19304, doi:10.1029/2009GL039795.

702 De Luca, G., R. Scarpa, L. Filippi, A. Gorini, S. Marcucci, P. Marsan, G. Milana and E.  
 703 Zambonelli (2000). A detailed analysis of two seismic sequences in Abruzzo, central  
 704 Apennines, Italy. *J. Seismol.* 4, 1–21.

705 Di Luzio, E., Mele, G., Tiberti, M.M., Cavinato, G.P., Parotto, M. (2009): Moho  
 706 deepening and shallow upper crustal delamination beneath the central Apennines.  
 707 *Earth. Planet. Sci. Lett.*, 280, 1-12, doi: 10.1016/j.epsl.2008.09.018.

708 Emergeo Working Group (2009). Rilievi geologici di terreno effettuati nell'area  
 709 epicentrale della sequenza sismica dell'Aquilano del 6 aprile 2009, report, Ist. Naz. di  
 710 Geofis. e Vulcanol., Rome. (Available at  
 711 <http://www.earthprints.org/handle/2122/5036>).

712 Hartzell, S. H., and T. H. Heaton (1983). Inversion of strong ground motion and  
 713 teleseismic waveform data for the fault rupture history of the 1979 Imperial Valley,  
 714 California, earthquake. *Bull. Seism. Soc. A.* **73**, 1553-1583.

715 Herrmann, R. B. (1973). Some aspects of band-pass filtering of surface waves, *Bull.*  
 716 *Seism. Soc. Am.* **63**, 703-711.

717 Herrmann, R. B., and C. J. Ammon (2002). Computer programs in seismology – 3.30:  
 718 surface waves, receiver functions and crustal structure,  
 719 [www.eas.slu.edu/People/RBHerrmann/CPS330.html](http://www.eas.slu.edu/People/RBHerrmann/CPS330.html)

720 Herrmann, R. B., and C. J. Ammon (2004). Computer programs in seismology – 3.30:

721 GSAC – Generic seismic application coding,  
 722 [www.eas.slu.edu/People/RBHerrmann/CPS330.html](http://www.eas.slu.edu/People/RBHerrmann/CPS330.html)  
 723 Herrmann, R. B., H. Benz and C. J. Ammon (2010). Monitoring the earthquake source  
 724 process in North America, *Bull. Seism. Soc. Am.* (submitted).  
 725 Hunstad, I., G. Selvaggi, N. D'Agostino, P. Englaand, P. Clarke, P. and M. Pierozzi  
 726 (2003). Geodetic strain in peninsular Italy between 1875 and 2001. *Geophys. Res.*  
 727 *Lett.* 30 (4).  
 728 Ligorria, J. P., and C. J. Ammon (1999). Iterative deconvolution and receiver-function  
 729 estimation, *Bull. Seism. Soc. Am.* 89, 1395-1400.  
 730  
 731 Lavecchia, G., F. Brozzetti, M. Barchi, J. Keller and M. Meinichetti (1994).  
 732 Seismotectonic zoning in east–central Italy deduced from the analysis of the Neogene  
 733 to present deformations and related stress fields. *Geol. Soc. Amer. Bull.* 106, 1107–  
 734 1120.  
 735 Mariucci, M. T., A. Amato, and P. Montone (1999). Recent tectonic evolution and  
 736 present stress in the northern Apennines (Italy), *Tectonics*, 18, 108– 118,  
 737 doi:10.1029/1998TC900019.  
 738 Massonnet, D., M. Rossi, C. Carmona, F. Adragna, G. Peltzer, K. Feigl, and T. Rabaute  
 739 (1993). The displacement field of the Landers earthquake mapped by radar  
 740 interferometry, *Nature*, 364, 138–142.  
 741 Michelini, A., L. Faenza, A. Lomax and M. Cocco 92 (2009). Appraisal of the  
 742 hypocentral location of the L'Aquila main shock, *EOS Trans. AGU*, 90(52), Fall.

743 Meet. Suppl., Abstract U23A-0027.

744 Michelini, A., L. Faenza, V. Lauciani, and L. Malagnini (2008). ShakeMap®  
 745 implementation in Italy, *Seism. Res. Lett.*, 79, 5, doi:10.1785/gssrl.79.5.688.

746 Montone, P., M. T. Mariucci, S. Pondrelli, and A. Amato (2004). An improved stress  
 747 map for Italy and surrounding regions (central Mediterranean). *J. Geophys. Res.* **109**,  
 748 B10410. doi:10.1029/2003JB002703.

749 Pace, B., L. Peruzza, G. Lavecchia, and P. Boncio (2006). Layered seismogenic source  
 750 model and probabilistic seismic-hazard analyses in Central Italy, *Bull. Seism. Soc.*  
 751 *Am.*, 96, 1, 107-132, doi:10.1785/0120040231.

752 Pondrelli, S., S. Salimbeni, A. Morelli, G. Ekström, M. Olivieri and E. Boschi (2010).  
 753 Seismic moment tensors of the April 2009, L'Aquila (Central Italy) earthquake  
 754 sequence, *Geophys. J. Int.* **180**, 238-242.

755 Scognamiglio, L., E. Tinti and A. Michelini (2009). Real-time determination of seismic  
 756 moment tensor for the Italian region, *Bull. Seism. Soc. Am.* **99**, 2223-2342.

757 Scognamiglio, L., E. Tinti and A. Michelini, D. Dreger, A. Cirella, M. Cocco, S. Mazza  
 758 and A. Piatanesi (2010). Fast determination of moment tensors and rupture history:  
 759 Application to the April 6<sup>th</sup> 2009, L'Aquila earthquake, *Seism. Res. Lett.* (in press).

760 Scrocca, D., C. Doglioni, P. Innocenti, A. Manetti, L. Mazzotti, L. Bertelli, L. Burbi and  
 761 S. D'Offizi (2003). *CROP Atlas: Deep Seismic Reflection Profile of the Italian Crust.*  
 762 *Mem. Descr. Carta Geol. Ital.* 62.

763 Selvaggi, G., B. Castello, and R. Azzara (1997). Spatial distribution of scalar seismic  
 764 moment release in Italy (1983-1996): seismotectonic implications for the Apennines,

765 Ann. Geofis., 40 (6), 1565-1578.  
 766 Stucchi, M. et al. (2007). DBMI04, il database delle osservazioni macrosismiche dei  
 767 terremoti Italiani utilizzate per la compilazione del catalogo parametrico CPTI04.  
 768 Quaderni di Geofisica, Vol. 49, pp.38  
 769 Tsai, Y., and K. Aki (1971). Amplitude spectra of surface waves from small earthquakes  
 770 and underground nuclear explosions, *J. Geophys. Res.* **76**, 3940-3952.  
 771 Walters, R.J., J.R. Elliot, N. D'Agostino, P.C. England, I. Hunstad, J.A. Jackson, B.  
 772 Parson, R.J. Phillips, and G. Roberts (2009). The 2009 l'Aquila earthquake (central  
 773 Italy): a source mechanism and implications for seismic hazard, *Geophys. Res. Lett.*,  
 774 36, L173312, doi:10.1029/2009GL039337.  
 775 Wessel, P. and W. Smith (1991). Free software helps map and display data, *Eos Trans.*  
 776 *AGU* **72** 441  
 777 Westaway, R. (1992). Revised nucleation point and fault rupture geometry for the 1980  
 778 Campania-Basilicata earthquake in southern Italy, *Geophys. J. Int.*, 109, 376-390.

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## 781 **Tables**

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Table 1.

Effect of frequency band on inversion results of the 2009/08/12 14:51:33 earthquake

Band (hz)	H (km)	Strike(°)	Dip (°)	Rake (°)	Mw	Fit
0.02 – 0.05	9	175	25	-55	3.34	0.47
0.02 – 0.10	7	185	20	-35	3.34	0.74
0.02 – 0.20	7	195	20	-25	3.22	0.66
0.02 – 0.40	7	200	25	-20	3.28	0.31
0.02 – 0.50	7	200	25	-15	3.27	0.20

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Table 2.

Source parameter variability

Parameter	Mean	Sigma
H (km)	0	0.9
Mw	0	0.05
Strike (°)	3	10
Dip (°)	-1	7
Rake (°)	1.3	10
Difference in P-axis (°)	10	10
Difference in T-axis (°)	7	7
Difference in B-axis (°)	11	11

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Table 3.

## Hypocenter parameters for the L'Aquila main shock

Date	Time (UTC)	Lat (°N)	Lon (°E)	H (km)	Source
2009/04//06	01:32:39	42.334	13.334	8.8	Initial ISIDE
2009/04//06	01:32:40.4	42.342	13.380	8.3	Final ISIDE
2009/04//06	01:32:39.7	42.341	13.371	13.7	<i>elocate</i> BB
2009/04//06	01:32:40.0	42.336	13.369	11.9	<i>elocate</i> ACCEL
2009/04//06	01:32:39.8	42.339	13.371	13.3	<i>elocate</i> BB+ACCEL
2009/04//06	01:32:40.8	42.347	13.380	9.5	Michellini et al (2009)
2009/04//06	01:32:40.7	42.350	13.376	9.3	Chiaraluce et al (2010)

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Table 4.

## Comparison of finite fault and point source inversions

Inversion	Frequency Band	Fit	Comment
Point source	0.01 – 0.025	0.700	STK=139, DIP= 55, RAKE=-94, Mw=6.13, H=5
	0.01 – 0.05	0.528	STK=138, DIP=56, RAKE=-97, Mw=6.03, H=5
Finite Fault	0.01 – 0.025	0.714	$\gamma = 0.0$ BB $V_r = V_s$
	0.01 – 0.05	0.642	$\gamma = 0.0$ BB $V_r = V_s$
	0.01 – 0.05	0.610	$\gamma = 0.0$ BB $V_r = 0.8V_s$
	0.01 – 0.05	0.542	$\gamma = 0.0$ BB $V_r = 0.6V_s$
	“	0.654	$\gamma = 0.0$ BB+ACCEL $V_r = V_s$
	“	0.648	$\gamma = 0.0$ BB+ACCEL $V_r = 0.8V_s$
	“	0.569	$\gamma = 0.0$ BB+ACCEL $V_r = 0.6V_s$
	“	0.646	$\gamma = 0.1$ BB+ACCEL $V_r = V_s$
	“	0.643	$\gamma = 0.1$ BB+ACCEL $V_r = 0.8V_s$
	“	0.553	$\gamma = 0.1$ BB+ACCEL $V_r = 0.6V_s$

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**Table A1**  
**Velocity models**

<b>H (km)</b>	<b>V<sub>P</sub> (km/s)</b>	<b>V<sub>S</sub> (km/s)</b>	<b>Density (kg/m<sup>3</sup>)</b>
<i>Initial model</i>			
1.5	5.0	2.86	2515
3	6.0	3.43	2687
3	6.0	3.43	2687
7	6.3	3.57	2754
15	6.0	3.43	2687
6	6.7	3.78	2850
8	7.1	3.99	2956
-	7.9	4.40	3212
<i>CIA (surface-wave)</i>			
1.5	3.75	2.14	2275
3	4.94	2.82	2485
3	6.01	3.43	2706
7	5.55	3.15	2609
15	5.88	3.36	2677
6	7.11	4.01	3010
8	7.10	3.99	3012
-	7.90	4.40	3276
<b>ACI (surface-wave and receiver function)</b>			
0.5	4.03	2.30	2323
0.5	3.81	2.18	2287
0.5	3.73	2.13	2271
1	4.54	2.59	2398
1	5.16	2.95	2532
1	5.58	3.18	2616
3	5.69	3.25	2637
3	5.38	3.05	2576
4	6.05	3.43	2714
5	5.51	3.15	2602
5	6.16	3.52	2747
5	5.76	3.29	2651
6	6.42	3.62	2828
8	7.35	4.13	3090
-	7.90	4.40	3276



## 802 Figure Captions

Figure 1. INGV catalog locations plotted to the nearest 0.1 degree for the time period October 1, 2008 through February 3, 2010 showing the locations of earthquakes, binned in the magnitude ranges 2.0 – 2.9, 3.0 -3.9, 4.0-4.9, 5.0-5.4 and 5.5-6.0. The symbol size is proportional to the magnitude bin. The numbers of earthquakes within each bin are 21239, 233, 25, 3 and 1, respectively. This study focuses on earthquakes with  $ML \geq 3.0$ .

Figure 2. Map showing earthquakes (triangles ) and stations (solid circles) used for the group velocity analysis to determine the regional velocity model. The dispersion paths sample the structure of Central Italy for a source in the central Apennines.

Figure 3. Comparison of observed (gray dots) and predicted (curves) Love- and Rayleigh-wave group velocity dispersion for different models: CIA, ACI, BAGH and TDMT. The observed groups velocities were obtained for the stations and earthquakes shown in Figure 2.

Figure 4. Left: comparison of the starting CIA and final ACI models for the joint inversion of surface-wave dispersion and receiver functions. Right: comparison of observed (solid) and ACI model predicted (dashed) P-wave receiver functions at the Mednet station AQU. The individual receiver functions are annotated on the right by event information.

Figure 5. Comparison of velocity models. The CIA model is used for source inversion.

Figure 6. Location of the earthquake and stations used to analyze the earthquake of

2009/08/12 14:51:33

Figure 7. Goodness of fit as a function of source depth for the earthquake of 2009/08/12 14:51:33. The best fitting mechanism at each source depth is plotted in a lower hemisphere projection. The best fit is for a depth of 7 km.

Figure 8. Comparison of observed (solid) and predicted (dashed) waveforms for the earthquake of 2009/08/12 14:51:33 as a function of absolute travel time. All traces represent ground velocity (m/s) filtered in the 0.02 – 0.1 Hz band. The peak amplitude is plotted to the left of each trace. The time shift of the synthetic with respect to the observed trace for the best waveform fit is given to the right of each trace. The station name is given to the right of the traces.

Figure 9. Comparison of inversion fits for station FDMO, at an epicentral distance of 78.7 km, for different frequency bands used for the inversion. The presentation is the same that of Figure 8.

Figure 10. Comparison of number of earthquakes in the catalog and the number of successful moment tensor solutions as function of INGV ML.

Figure 11. (a) Comparison of  $M_w$  from moment tensor inversion to INGV automatic determination of ML; (b) Comparison of moment tensor depths to INGV automatic location depths ; (c ) Comparison of moment tensor depths to 1-D relocations of Chiaraluce et al (2010).

Figure 12. Moment tensor solutions for the L'Aquila sequence shown in a lower hemisphere equal-area projection. The colors indicate the source depth determined

by broadband modeling. Note that the main shock (largest event) depth is not consistent with the depths of neighboring aftershocks. Subsequent relocations place it about 3 km east, where it is still slightly shallow compared to aftershocks.

Figure 13. Goodness of fit as a function of source depth for the L'Aquila main shock using the 0.01 – 0.05 Hz band for inversion. The best fit is at 29 km, the limit of the depth search, although there is a local maximum at a depth of 5km.

Figure 14. Goodness of fit as a function of source depth for the L'Aquila main shock using the 0.01 – 0.025 Hz band for inversion. The best fit is at a depth of 5 km.

Figure 15. Locations of broadband stations (solid circles) and accelerometers (inverted triangles) used for the analysis of the main shock (upright triangle) which is indicated by the triangle.

Figure 16. Comparison of observed (solid) and predicted (dashed) waveforms as a function of travel time for the best fit point source solution using the 0.01 – 0.025 Hz frequency band. The figure annotation is as for Figure 8. Note the large positive time shifts of the synthetic with respect to the observed waveform and also the high frequency motions on parts of the predicted surface-wave arrival.

Figure 17. Comparison of finite fault waveforms (dashed) to observed ground velocities (solid) in the 0.01 – 0.025 Hz band. No spatial smoothing is assumed and rupture velocity equals the local S-wave velocity. The misalignment of the surface-wave arrival at larger distances indicates the need for slight changes in the velocity model.

Figure 18. Location of finite fault subevents with respect to our moment tensor solutions.

Shaded circles – events for which moment tensor inversions were determined in this study with the shading a function of source depth; the largest circle is the location of the initial automatic solution for the main shock. Star – initiation point for finite fault rupture. Diamonds – finite fault sub-events. Small squares indicate nearby cities: LA – L'Aquila, PA – Paganica and PP – Poggio Picenze. The size of all events is scaled with magnitude.

Figure 19. Sensitivity of finite fault inversion of broadband data in the 0.01 – 0.025 Hz band to smoothing for a fixed rupture velocity equal to local S-wave velocity. a) smoothing parameter = 0.0, b) smoothing parameter = 1.0 and c) smoothing parameter = 0.1; The rupture velocity was set to the local shear-wave velocity. The solid circle indicates the hypocenter and the diamond the point of maximum slip. The dashed gray lines indicate the rupture timing in seconds. The slip contours increase from 25 to 700 cm with the same shading for all images.

Figure 20. Sensitivity of finite fault inversion of broadband data in the 0.01 – 0.05 Hz band to rupture velocity. A fixed smoothing parameter of 1.0 is used. The left column data set consists of only the regional broadband data, with the right column data set adds local acceleration records. Rupture velocity decreases as a function of the shear-wave velocity from top to bottom as 1.0, 0.8 and 0.6. The solid circle indicates the hypocenter and the diamond the location of maximum slip. The dashed gray lines indicate the rupture timing in seconds. The slip contours increase from 25 to 700 cm with the same shading for all images.

Figure 21. Sensitivity of finite fault inversion of broadband data in the 0.01 – 0.05 Hz band to rupture velocity. A fixed smoothing parameter of 0.1 is used. The left column data set consists of only the regional broadband data, with the right column data set adds local acceleration records. Rupture velocity decreases as a function of the shear-wave velocity from top to bottom as 1.0, 0.8 and 0.6. The solid circle indicates the hypocenter and the diamond the location of maximum slip. The dashed gray lines indicate the rupture timing in seconds. The slip contours increase from 25 to 700 cm with the same shading for all images.

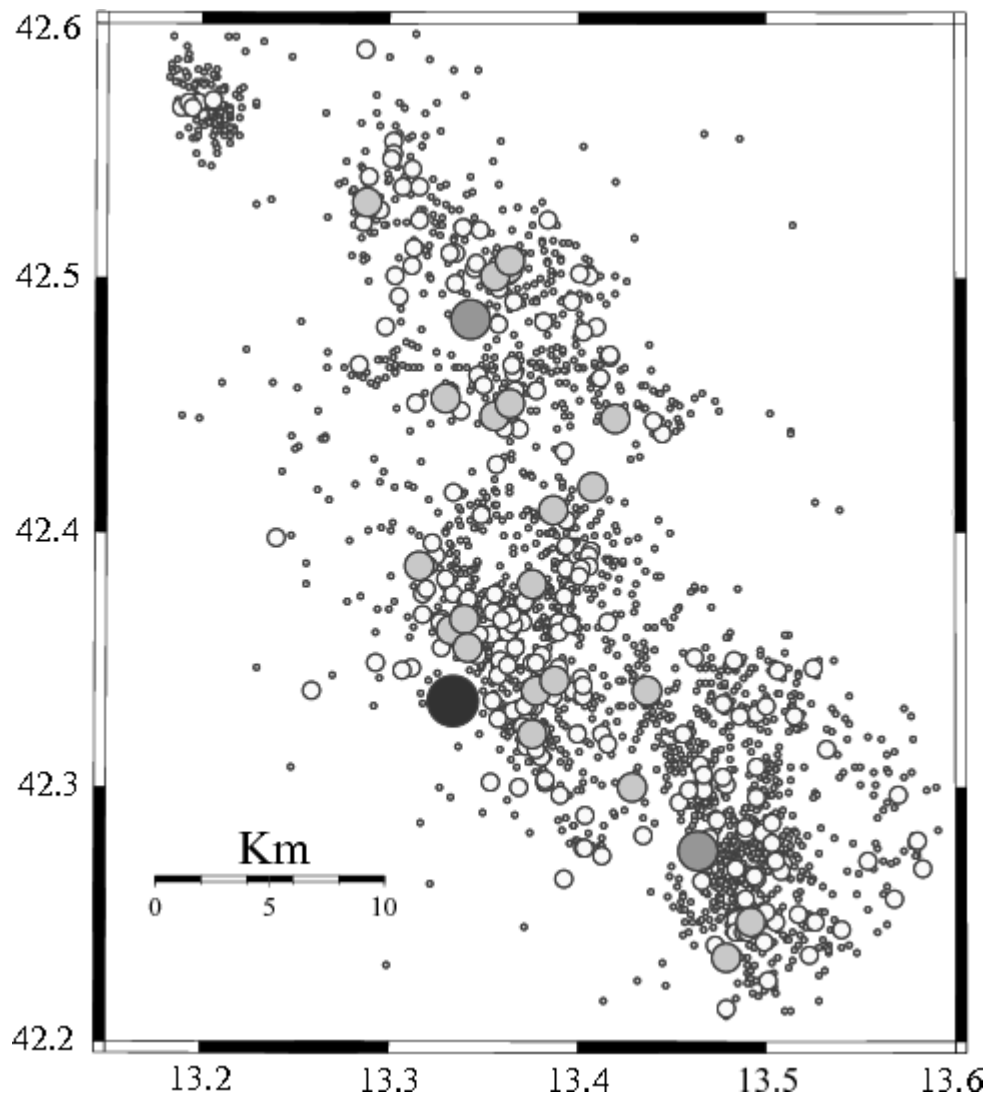


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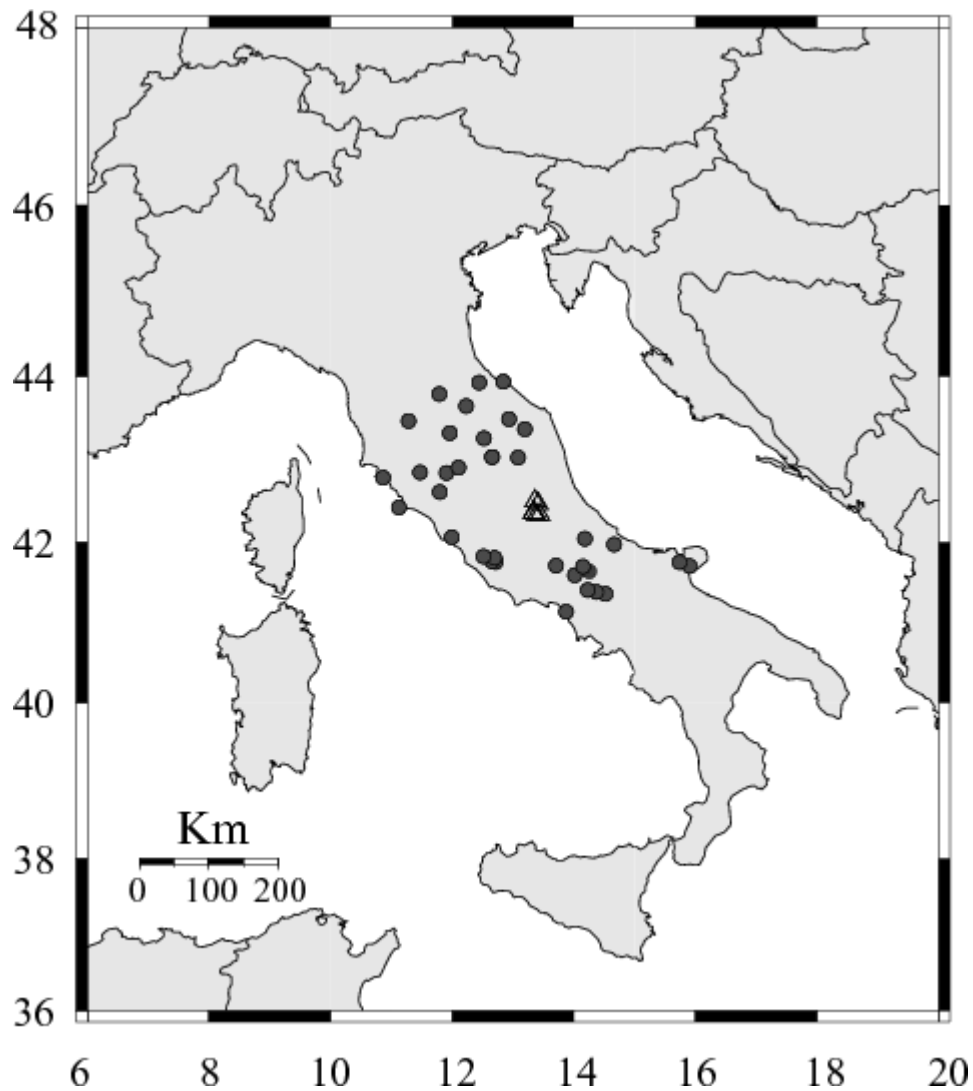


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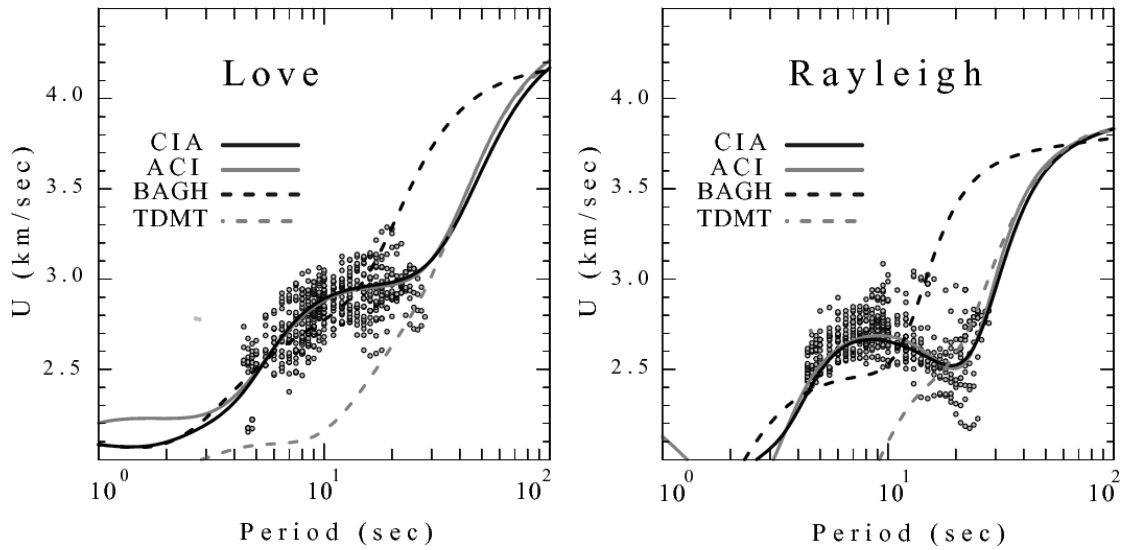
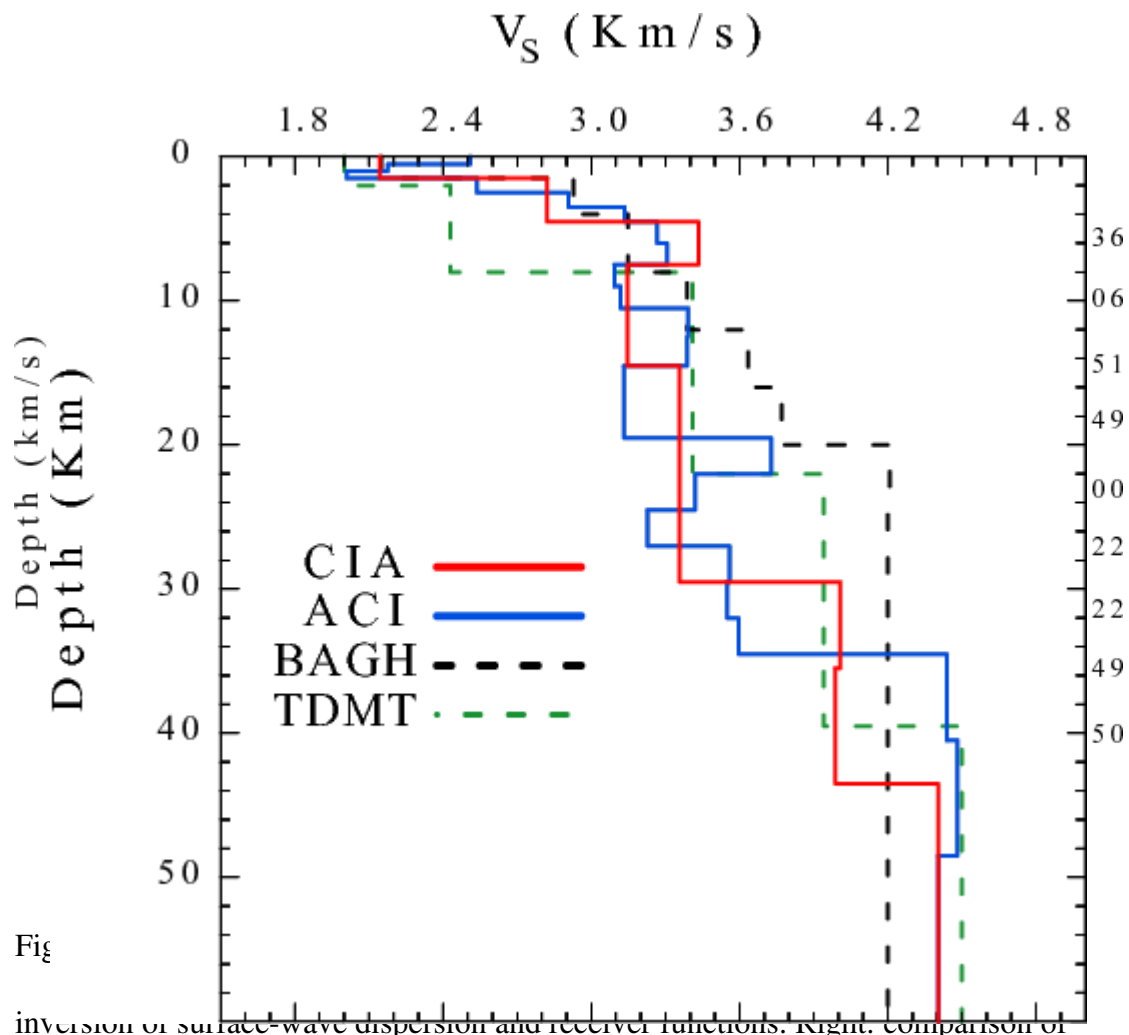


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inversion of surface-wave dispersion and receiver functions. Right. Comparison of

observed (black) and ACI model predicted (gray) P-wave receiver functions at the Mednet station AQU. The individual receiver functions are annotated on the right by event information and on the left by station name, Gaussian filter parameter, reduction of variance and ray parameter (sec/km).

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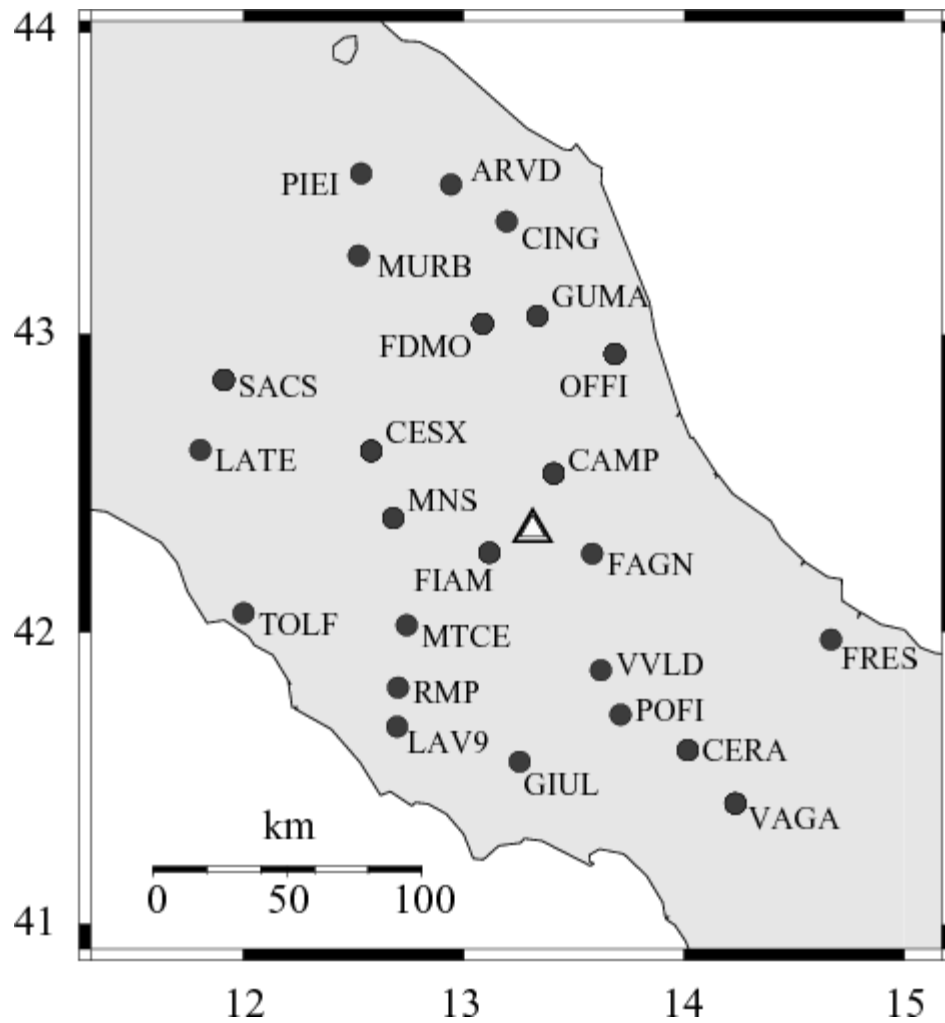


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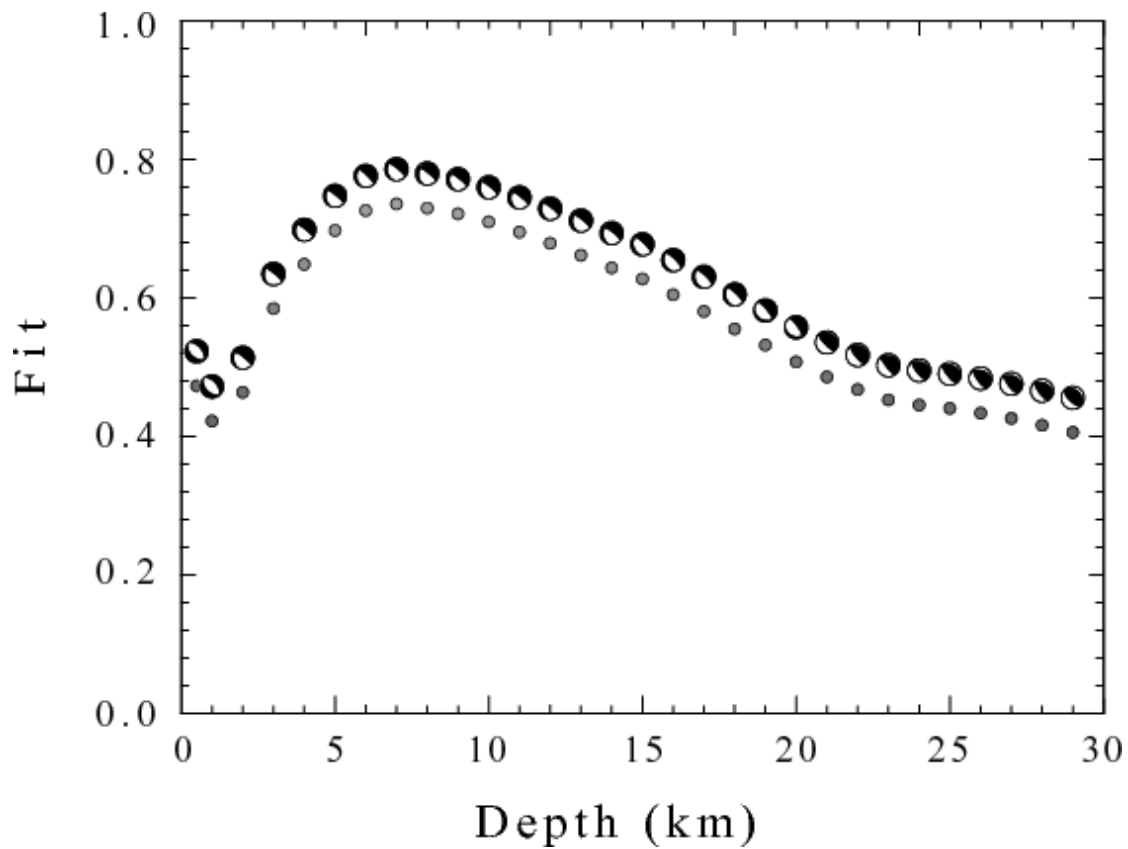


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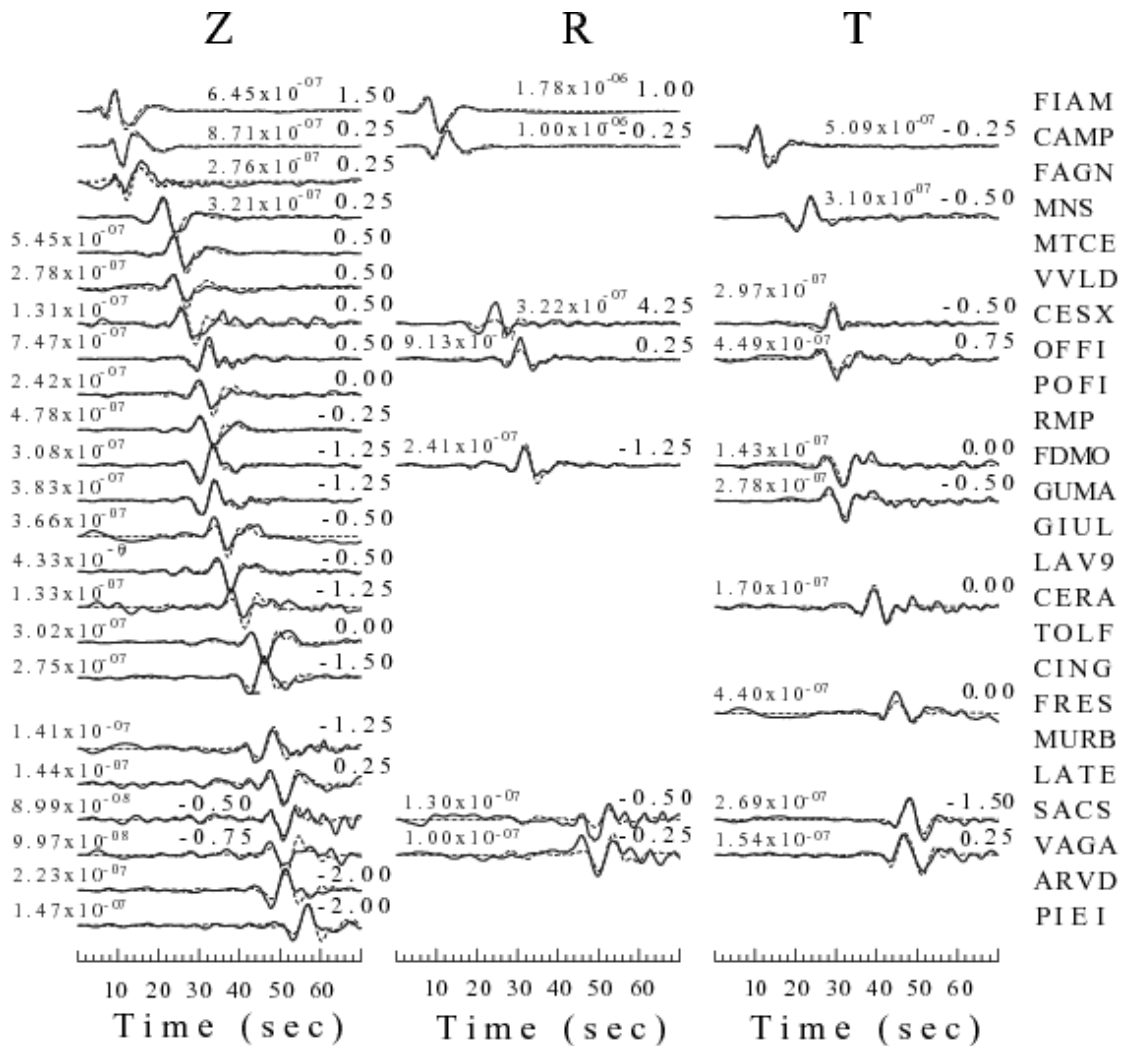


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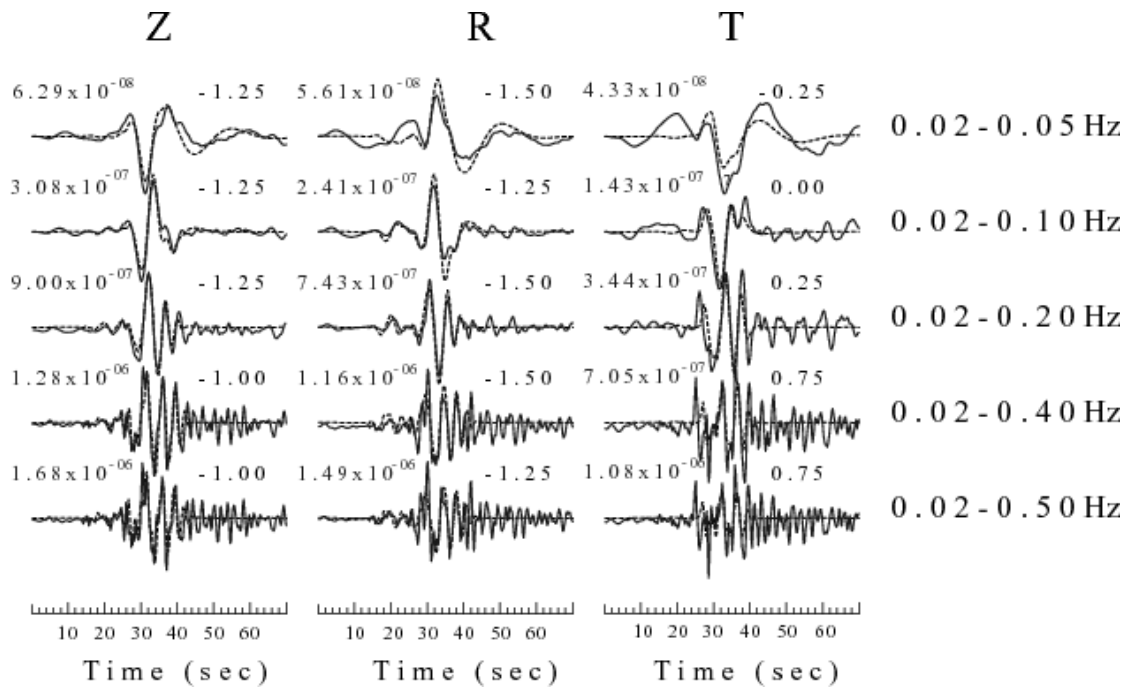


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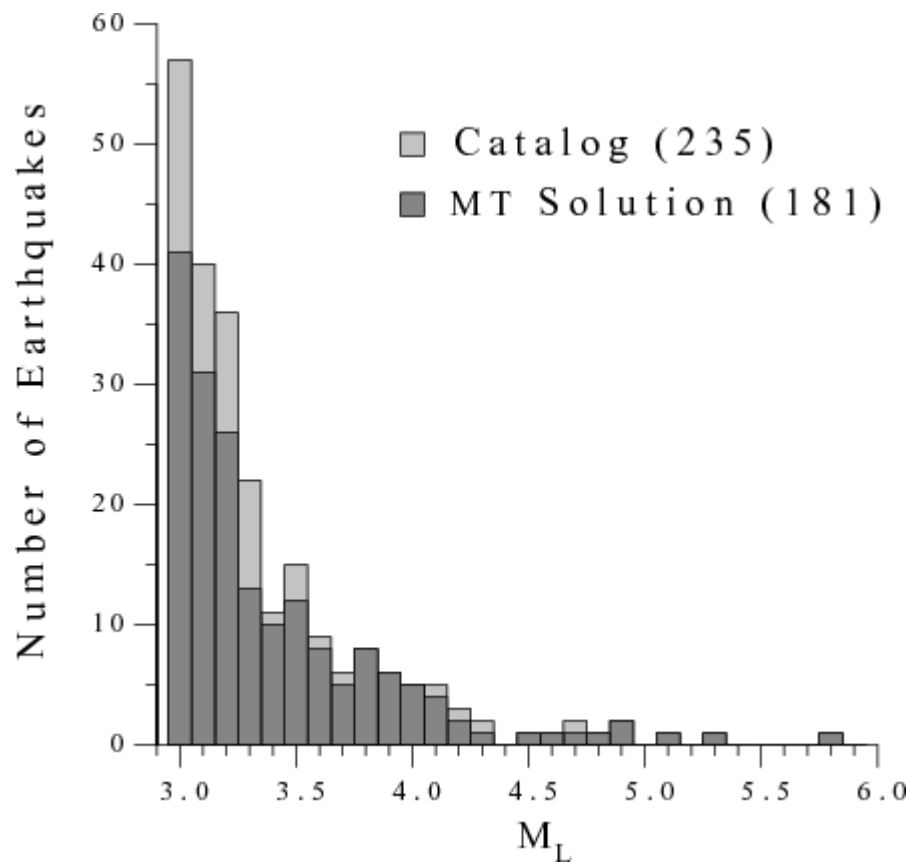


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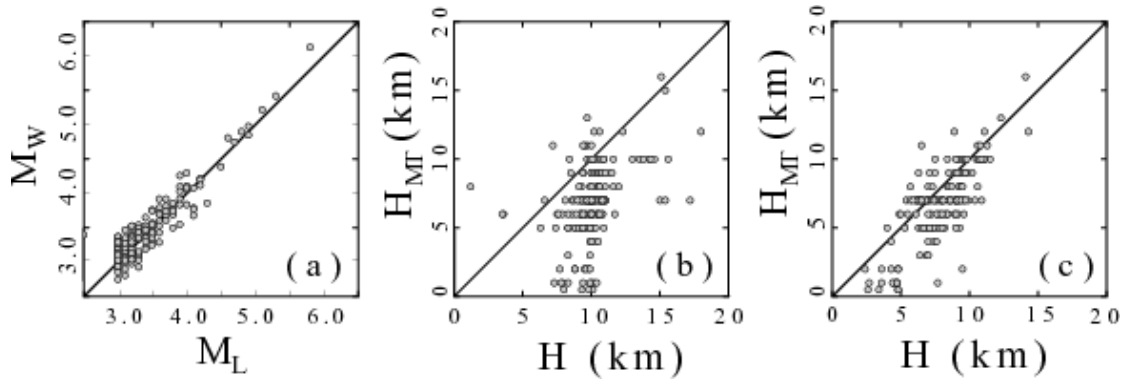


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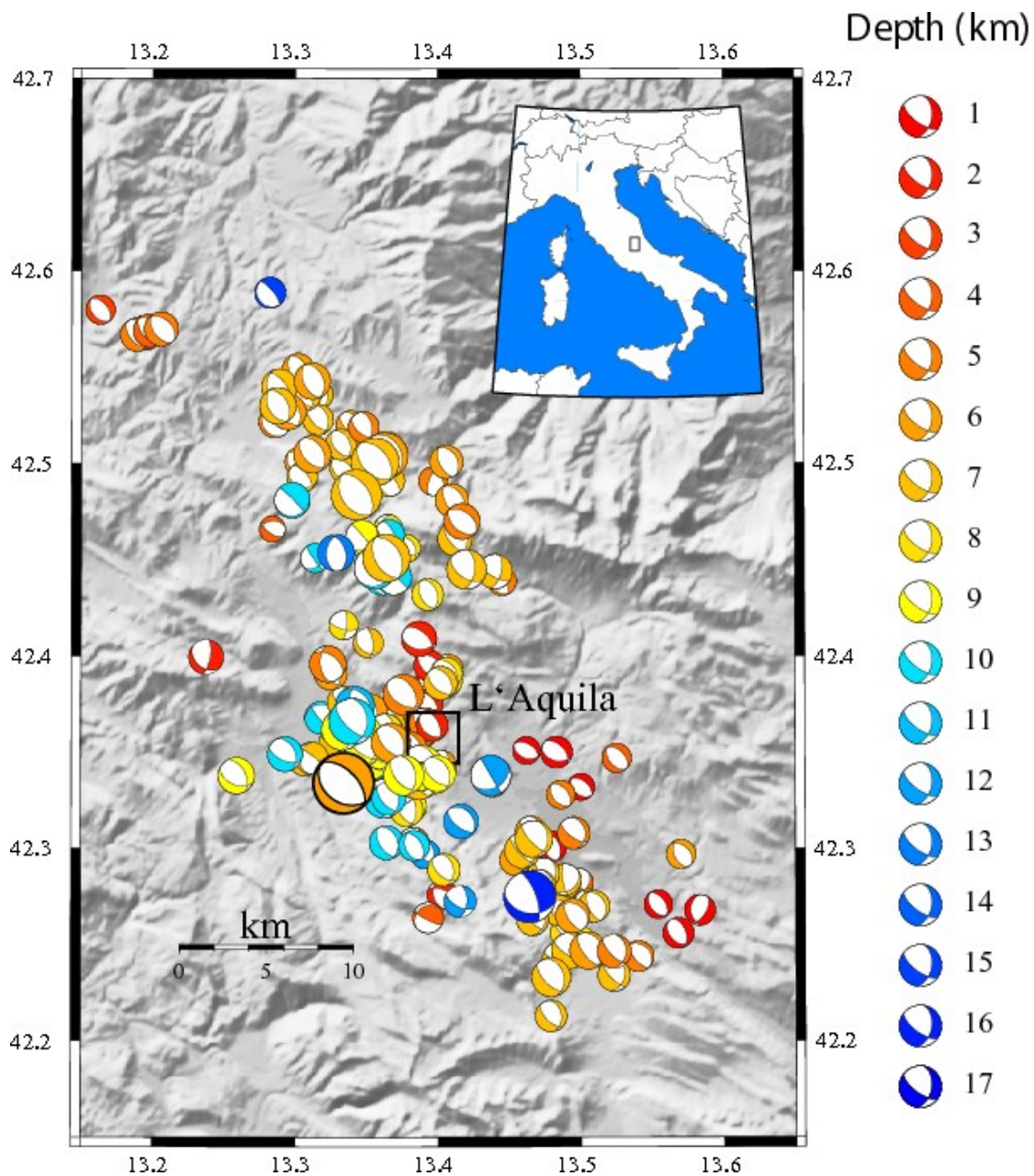


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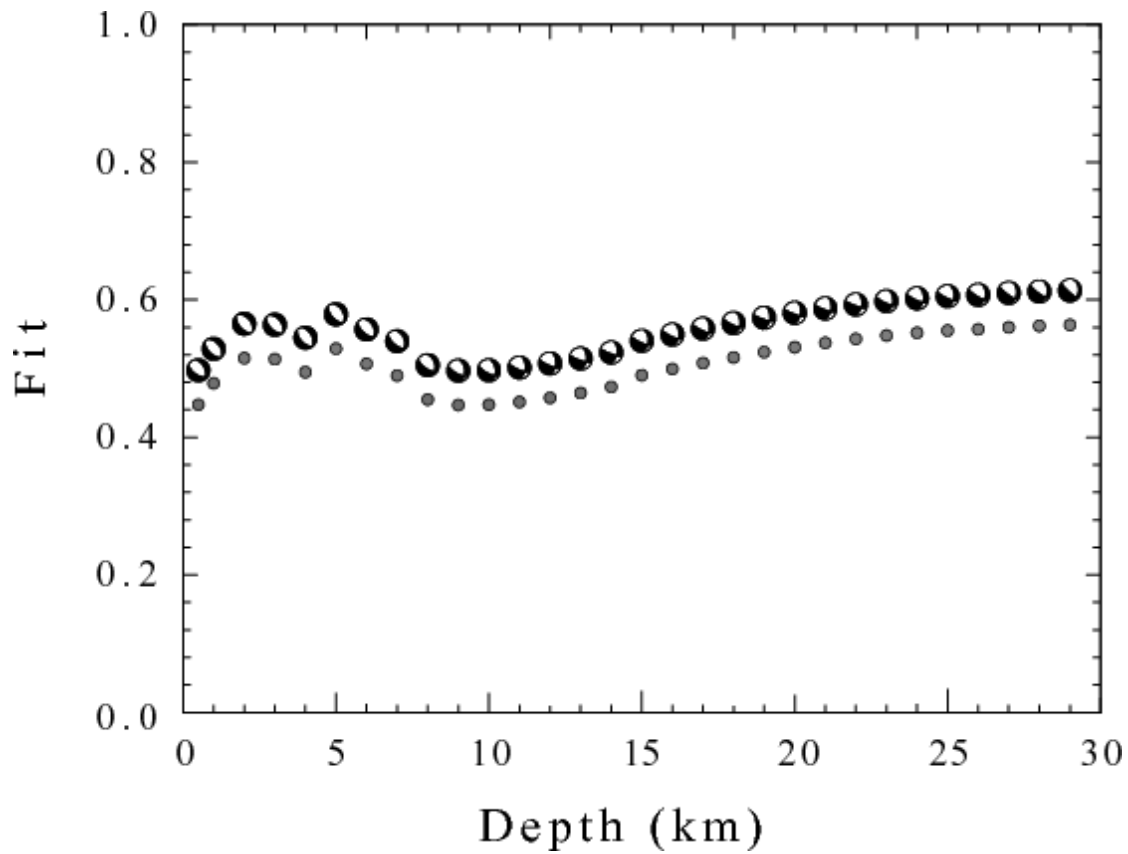


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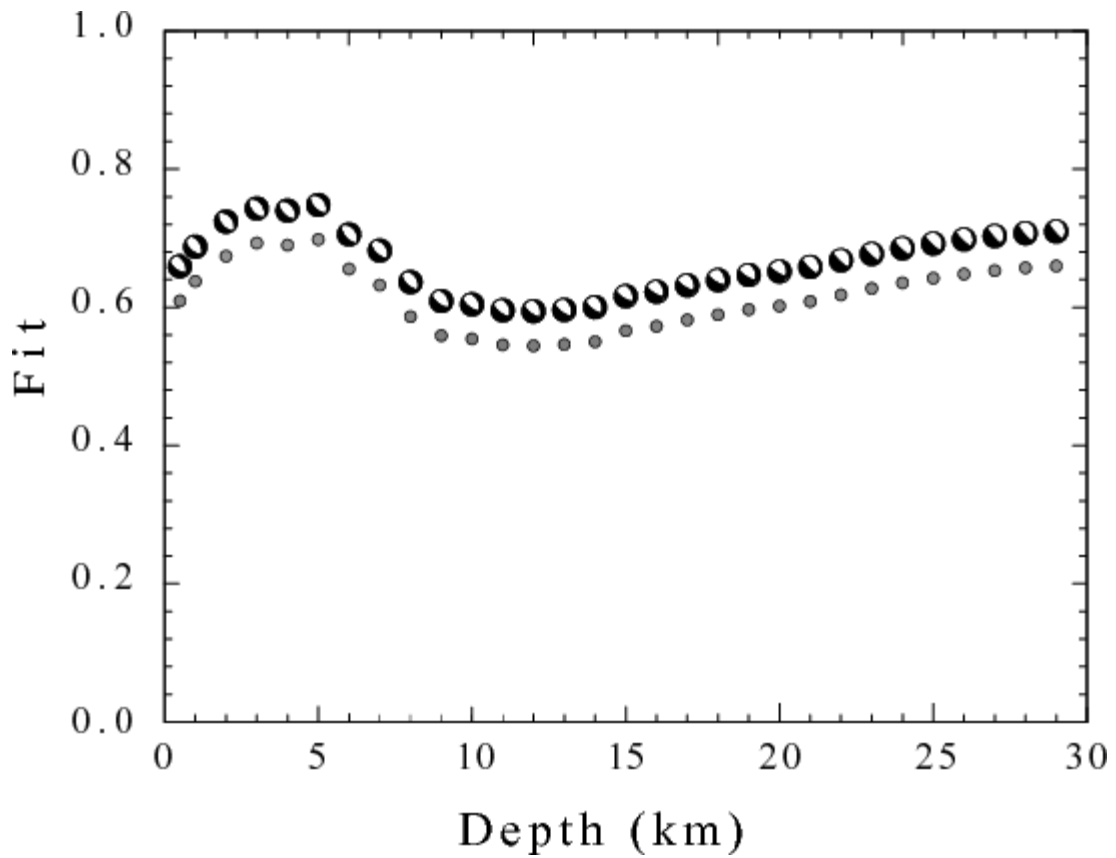


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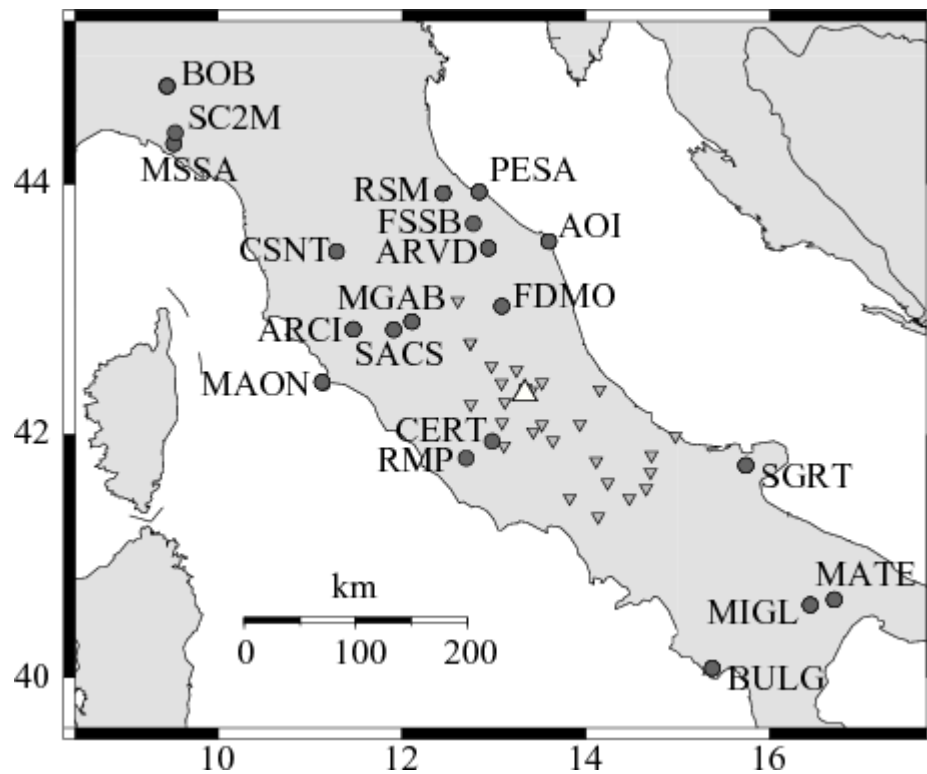


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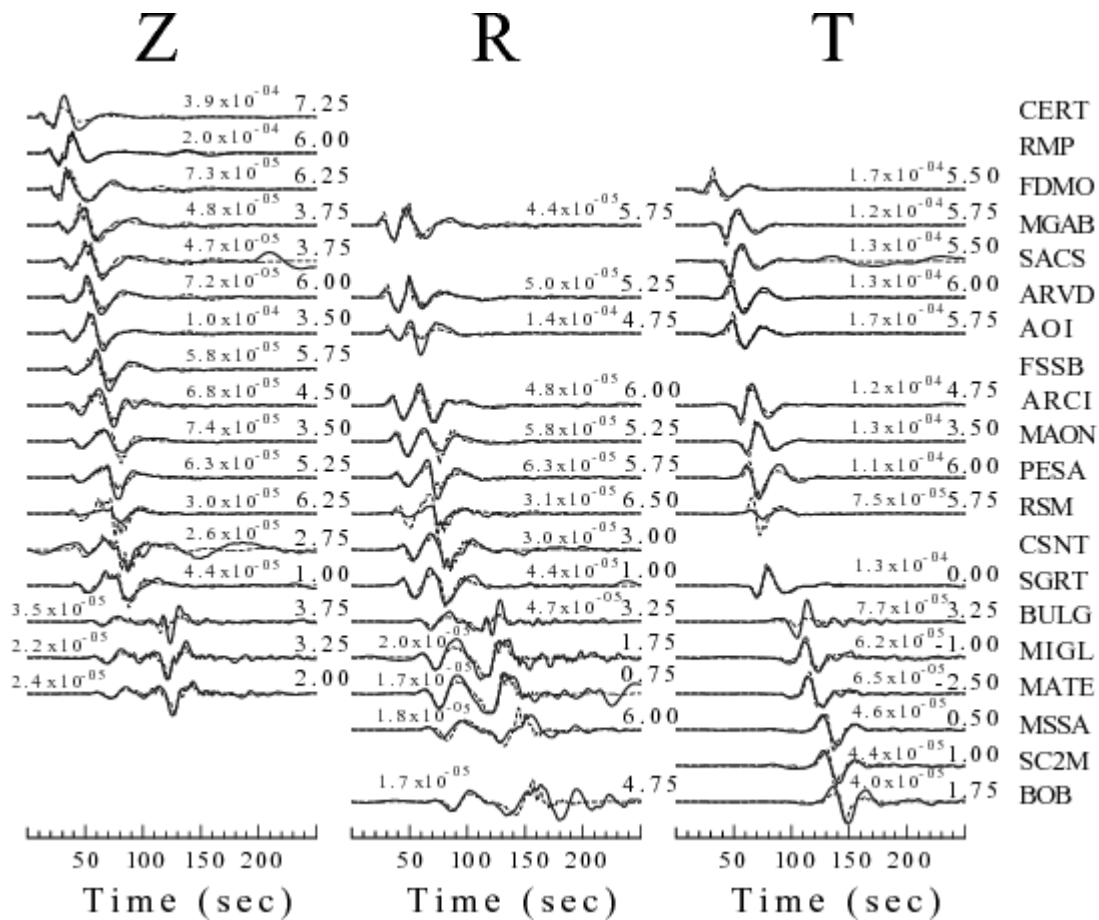


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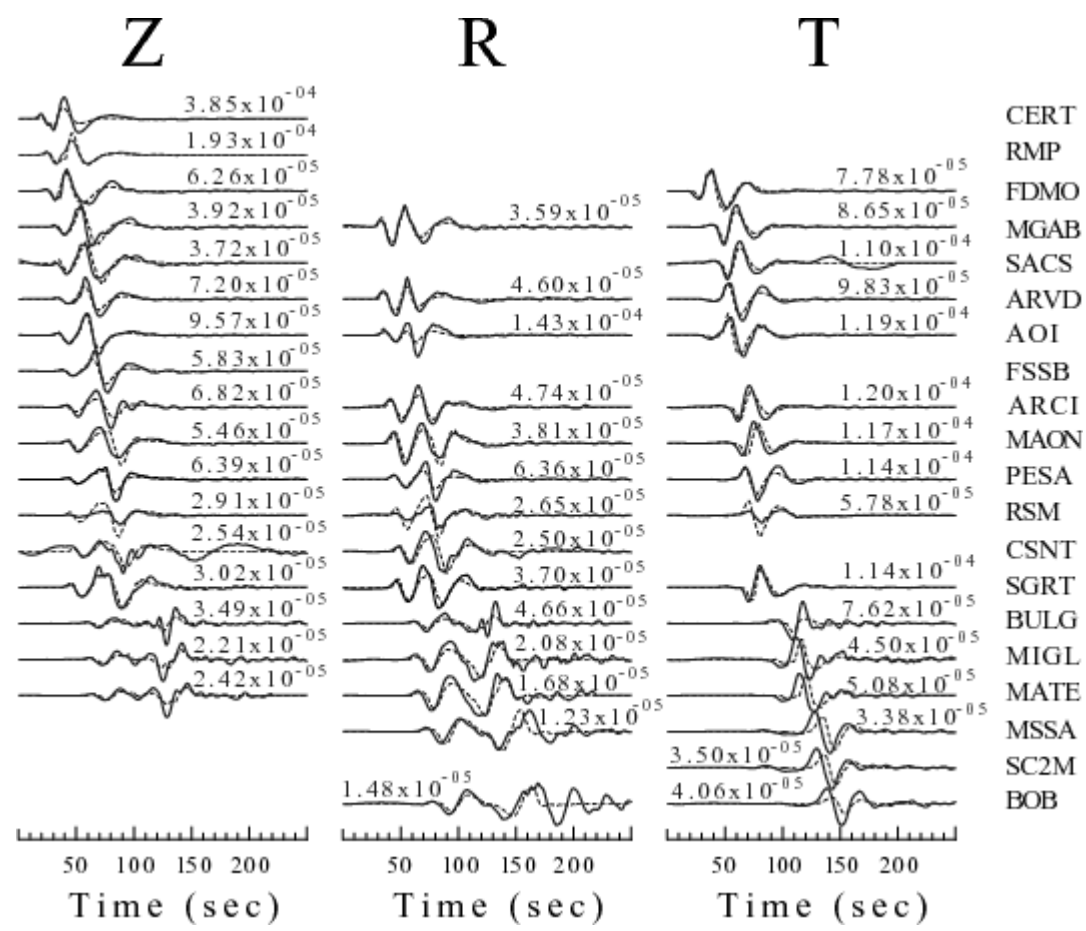


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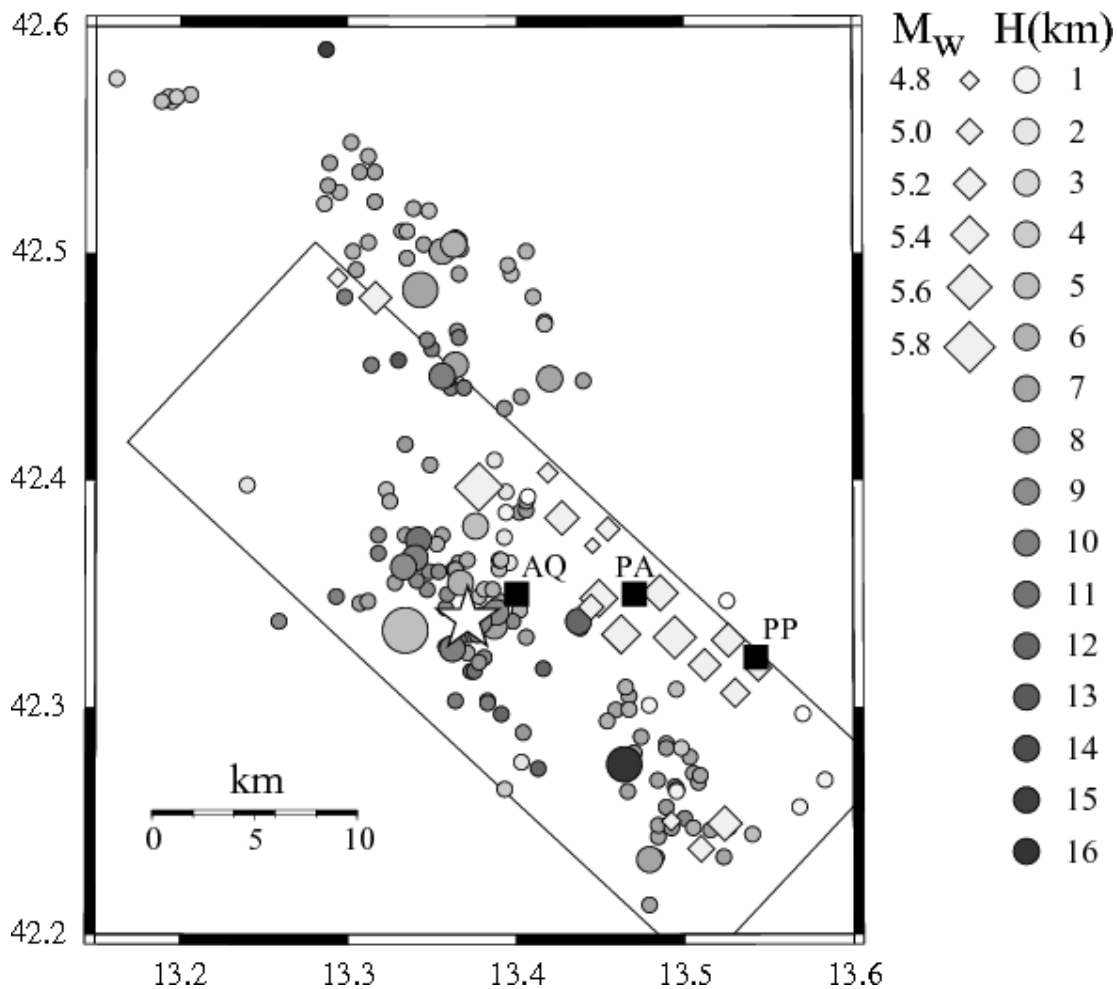


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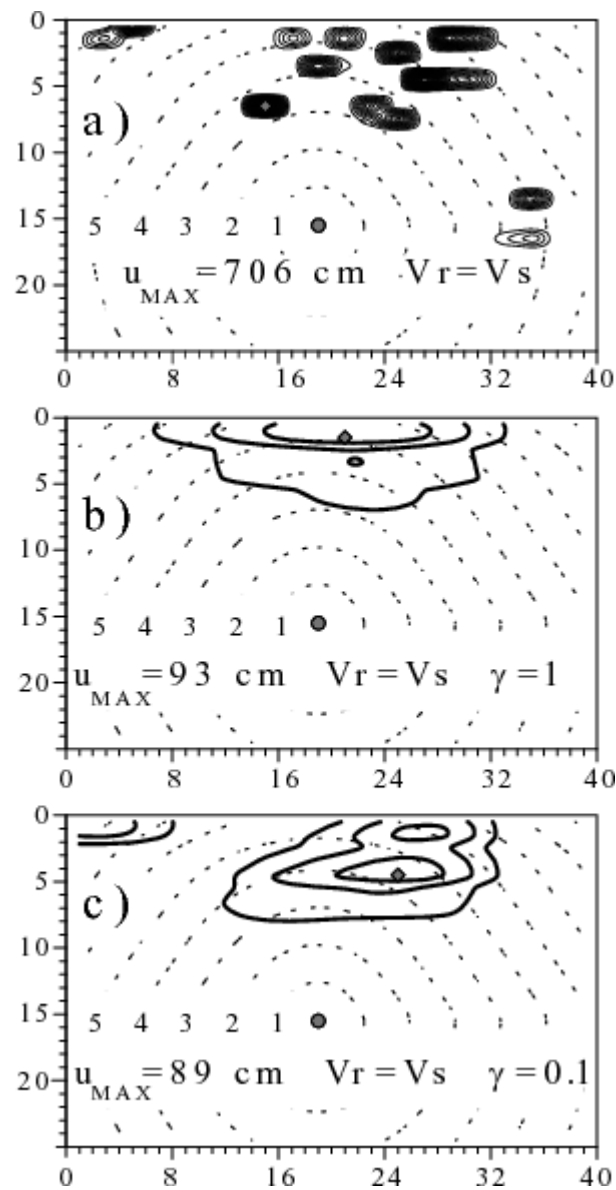


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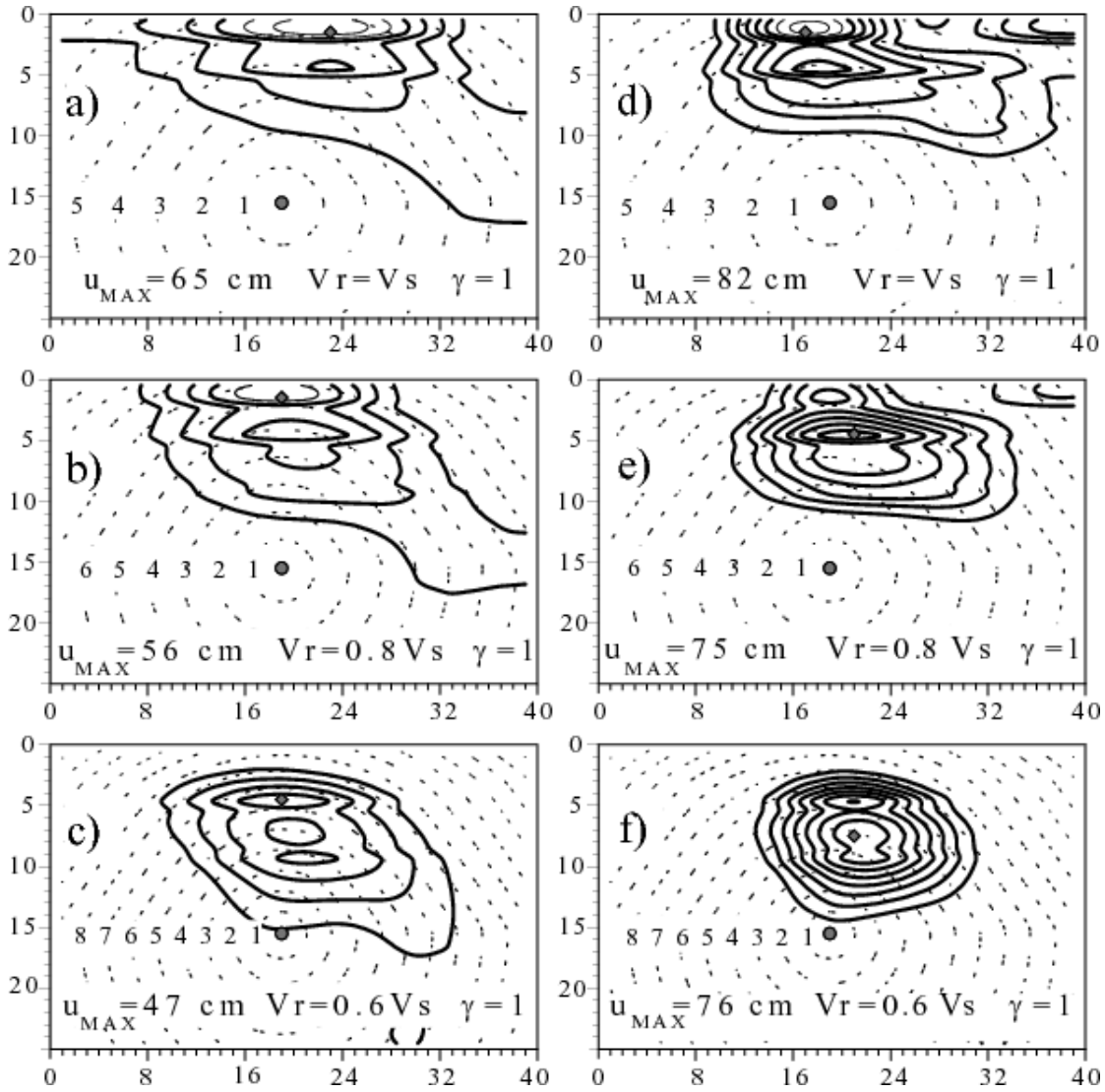


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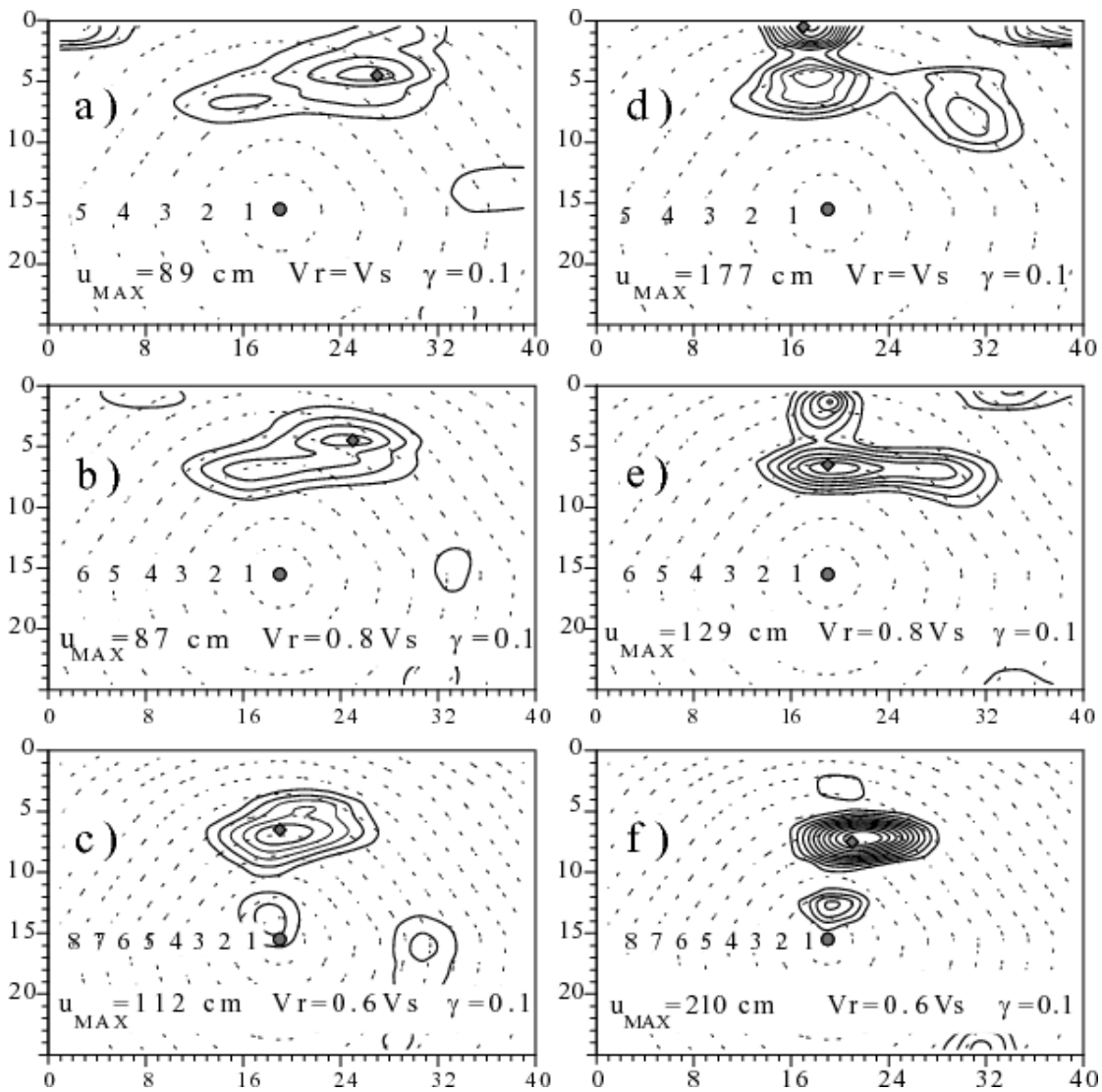


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### **Electronic Supplement**

876 This can be viewed at

877 [http://www.eas.slu.edu/Earthquake\\_Center/MECH.IT/Herrmann\\_Malagnini\\_Munafò\\_Suppl.html](http://www.eas.slu.edu/Earthquake_Center/MECH.IT/Herrmann_Malagnini_Munafò_Suppl.html)

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